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Geological Evolution and Analysis of Confirmed or Suspected Gas Hydrate Localities

Volume 3. Basin Analysis, Formation and Stability of Gas Hydrates in the Western Gulf of Mexico

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PREFACE

This document is Volume 3 of a series of reports entitled "Geological Evolution and Analysis of Confirmed or Suspected Gas Hydrate Localities." Volume 3 is an analysis of the "Formation and Stability of Gas Hydrates in the Western Gulf of Mexico." This report presents a geological description of the western Gulf of Mexico, including regional and local structural settings, geomorphology, geological history, stratigraphy, and physical properties. It provides the necessary regional and geological background for more in-depth research of the area. Detailed discussion of bottom simulating acoustic reflectors, sediment acoustic properties, and distribution of hydrates within sediments is also included. The formation and stabilization of gas hydrates in sediments are considered in terms of phase relations, nucleation, and crystallization constraints, gas solubility, pore fluid chemistry, inorganic diagenesis, and sediment organic content. Together with a depositional analysis of the area, this report is a better understanding of the thermal evolution of the locality. It should lead to an assessment of the potential for thermogenic hydrocarbon generation.

Project Manager
Gas Hydrates

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BASIN ANALYSIS, FORMATION AND STABILITY OF GAS HYDRATES IN THE WESTERN GULF OF MEXICO

By Jan Krason, Patrick Finley, and Bernard Rudloff

EXECUTIVE SUMMARY

The geological relationships affecting gas hydrate formation and stability in the western Gulf of Mexico were evaluated by thorough basin analysis of the study region. Particular emphasis was placed on examining the factors which were determined in previous studies to be critical to offshore gas hydrate formation.

This work was performed for the U.S. Department of Energy - Morgantown Energy Technology Center by Geoexplorers International as part of a worldwide study of the geological characteristics of 24 offshore locations with confirmed or inferred presence of gas hydrates. Results of this study are summarized in Table 1.

The study region comprises the entire Gulf of Mexico west of 88° west longitude, an area of approximately 940,000 km^2. Nearly 20% of the study region is underlain by water of insufficient depth to stabilize gas hydrates. This huge region is subdivided into three general geologic provinces, the deep central Gulf of Mexico, the northwestern margin of the Gulf of Mexico, and the western martin of the Gulf of Mexico.

The geology of the Gulf of Mexico has been studied in detail by many industrial and research organizations. The geological evolution presented here is a consensus distilled from the voluminous published and unpublished literature available on the region. Proprietary industry data were not included except when no comparable information was available in the public domain. Over 40,000 km of seismic data were examined for bottom simulating reflectors and to refine geological interpretations.

The deep central Gulf of Mexico is floored by crust of debatable origin. Isolated Triassic red-bed sequences on the crust are overlain by extensive Jurassic evaporite deposits. Subsidence and return of deep marine conditions led to the deposition of deep marine clastics and foraminiferal ooze. Increased sediment influx in the Tertiary and Quaternary, and glacially induced sea level fluctuations resulted in an increase in sedimentation rate and a shift to turbidite deposition on the abyssal plain.

The geology of the northwestern margin of the Gulf of Mexico is dominated by salt diapirism and heavy Cenozoic sedimentation. Cretaceous carbonate platforms were covered with thick sediments from Tertiary orogeny In response to this sediment loading, which was

accelerated in the Quaternary, vertical and lateral flow of salt was initiated, which deformed the sediments and resulted in the characteristic structure and sea floor topography of the area. Pleistocene sedimentation from the Mississippi River further blanketed the continental slope and produced a large alluvial fan.

The western margin of the Gulf of Mexico is characterized by gravity induced detachment folds along the Mexican continental slope and salt diapirs in the south. Sliding along a shale decollement zone produced large, regular north trending folds on the continental slope offshore of northwestern Mexico. Southeast of these folds, a large area of diapiric structures, the Campeche Knolls document salt mobilization in the absence of abnormal Quaternary sediment loading.

Thermogenic gas hydrates were recovered from the northwestern margin of the Gulf of Mexico. Indirect evidence suggests that thermogenic gas hydrates occur on diapirs in the Campeche Knolls. Burial history reconstructions indicate that hydrocarbons in these gas hydrate deposits must have migrated at least 2,000 to 3,000 m vertically. Their association with salt structures appears to be related to the structural migrational pathways that diapirism provided. The presence of ethane through butane in the migrated thermogenic gas stabilizes these gas hydrates relative to methane hydrates, but saline pore waters due to solution of diapirs destabilize thermogenic gas hydrates.

Biogenic gas hydrates were recovered from the northwestern margin, and are inferred to have been drilled in the deep central Gulf of Mexico. Sedimentary organic matter appears to have been preserved by rapid sedimentation throughout the region. Total organic carbon content of the sediments varies from 0.2% to 3%. Large amounts of biogenic methane were generated in the deep central Gulf area from sediments with less than the accepted lower threshold of organic carbon content for microbial methanogenesis. Biogenic gas hydrates occur in a wide range of lithologies, and may show an association with volcanic detritus in host sediments. Probable gas hydrate locations in the Gulf of Mexico generally do not display the decrease in pore water salinity with depth which is often associated with gas hydrates.

Bottom simulating reflectors (BSRs) covering approximately 5,000 km^2 abound in anticlines offshore of the Mexican coast between Tampico and Veracruz. The BSRs are found in water depths of 1,200 to 2,700 m and at 400 - 600 m subbottom. Dense spacing of seismic lines permits determination of the areal extent of some BSRs and the structural closure beneath these BSRs. Some BSRs can be traced between as many as six seismic sections over distances of up to 80 km. Structural positions of the BSRs are consistent with the interpretation of free gas beneath the hydrate zone increasing the amplitude of the reflection. Thus it is likely that gas hydrates exist in adjacent areas where no BSRs are found due to a lack of underlying free gas.

Given the very large data gaps on areal extent, vertical distribution, and degree of pore occupancy of hydrates in the Gulf of Mexico, estimates of gas contained in hydrates and as free gas beneath hydrates are speculative. However, reasonable assumptions of these parameters permits very rough estimates of in-place gas volumes at standard conditions:

Area Contained in Hydrates	Range of Estimates (TCF)	Most Probable Estimate (TCF)
Deep Central Gulf	5 - 3,000	1,000
Northwestern Margin	1 - 2,000	200
Western Margin	12 - 4,000	1,200

Gas possibly trapped beneath gas hydrates is calculated only for the 5,000 km^2 area covered by BSRs. Estimates range from 7 to 350 TCF with a most probable figure of 100 TCF.

Table 1. SUMMARY DATA OF BASIN ANALYSIS, FORMATION AND STABILITY OF GAS HYDRATES IN THE WESTERN GULF OF MEXICO

FACTORS / BASIN / SUB-BASIN	CENTRAL GULF BASIN	NORTHWESTERN CONTINENTAL MARGIN	SOUTHWESTERN CONTINENTAL MARGIN
	Sigsbee Abyssal Plain Mississippi Fan	Offshore Texas and Louisiana	Mexican Ridges, Campeche Knolls
BASIN ANALYSIS			
Location			
Longitude: latitude	88° - 95°W; 22° - 26°N	88° - 97°W; 26° - 29°N	94° - 97°W; 19° - 26°N
Areal extent, km^2	240,000 km^2	190,000 km^2	200,000 km^2
Geomorphology	Abyssal Plain, deep sea fan	Continental slope	Continental slope
Geomorphologic sub-unit		Intraslope basins, submarine canyons	Anticlinal ridges, salt knolls
Geology	Moderately well defined	Very well defined	Moderately well defined
Structural setting	Quiescent oceanic basement	Extensive salt diapirism, growth faulting	Gravity induced folds, salt and shale diapirism
Stratigraphy	Triassic through Quaternary	Triassic through Quaternary	Triassic through Quaternary
Lithology	Salt, shales, turbidites	Salt, deltaic deposits	Salt, shales, turbidites
Sedimentary environments	Pelagic, hemipelagic	Deltaic, hemipelagic	Hemipelagic, pelagic
Sediment source	North American craton via Mississippi River	North American craton via Mississippi River	East-central Mexican highland
Rate of sedimentation	Average 5 - 10 cm/1000 yrs.	Average 20 - 200 cm/1000 yrs.	Average 5 - 10 cm/1000 yrs.
Sediment flux, $mg/cm^2/yr$	2 - 350, average 5	No data	Average 4
Organic matter flux, $mg/cm^2/yr$	Average 1	No data	Average 1
Geochemistry	Poorly known	Well known	Poorly known
Total organic matter content, weight %	0% - 5%, average 0.4%	Average 0.4%	Average 0.4%
Source of organic matter	Terrestrial, marine	Terrestrial	Terrestrial
Preservation of organic matter	Well preserved	Well preserved	Well preserved
Depth of thermal maturity, m	2,000 m	3,000 m	2,000 m
Geochemical anomalies	No data	No data	No data
Sediment alteration	No data	No data	No data
Physical and geophysical features	Well defined	Well defined	Well defined
Thickness, m	12,000 m	14,000 m	8,000 m
Porosity	70 - 20%	70 - 20%	70 - 20%
Permeability, md (millidarcy)	No data	No data	No data
Geothermal gradient, °C/m	4°C/100 m	2 - 8°C/100 m	3 - 5°C/100 m
Heat flow, HFU (heat flow unit)	Average 0.83	Average 0.7	Average 1.2
GAS HYDRATES FORMATION AND STABILITY			
Direct evidence	None	Recovery of hydrates	None
Type of gas hydrate occurrence	Unknown	Disseminated to nodular, thermogenic and biogenic	Unknown
Indirect evidence	Anomalous core degassing	Gas seeps	Anomalous core degassing
Bottom simulating reflector(s), BSR	None	None	Abundant
Areal extent of the bottom simulating reflector(s), km^2	N/A	N/A	8,000 km^2
Quality of seismic data	Good	Good	Good
Inferred evidence	Temperature, pressure	Temperature, pressure	Temperature, pressure
Location	24.8°N, 95.1°W; 93.3°W, 23.7°N	26°-28.5°N; 90°-92°W	19°-23°N; 93°-97°W
Sea water depth, m	3,000 - 3,700 m	400 - 3,000 m	400 - 3,000 m
Sub-sea bottom depth, m	100 - 800 m	0 - 50 m	100 - 600 m
Hydrostatic pressure at sea floor, atmosphere	300 - 380 atmosphere	42 - 320 atmosphere	42 - 320 atmosphere
Temperature at sea floor, °C	4°C	5 - 15°C	5 - 12°C
Gas hydrates host formation	Turbidites	Hemipelagites	Turbidites (?)
Age of gas hydrates host formation	Miocene - Pliocene	Pleistocene	Miocene - Pleistocene
Gas hydrates stability zone	900 m	200 m	600 m
Initial porosity of gas hydrates host formation, vol. %	70%	70%	70%
Isotopic composition of gas, δ^{13}C ‰	-78 to -84‰ biogenic ?	-45‰ thermogenic -74‰ biogenic	-55‰ thermogenic no data biogenic
Pore water salinity, ‰ at depth m	3.5‰ at 100 m	Increase with depth near diapirs	Increase with depth near diapirs
Associated hydrocarbons	None	C_2 - C_5, biodegraded petroleum	C_2 - C_4, Site 88
Time of gas hydrates stabilization	No data	No data	No data
Source of gas hydrates	Biogenic	Thermogenic and biogenic	Thermogenic and biogenic
Evidence for free gas under gas hydrate zone	None	None	BSRs on structural highs
Estimated/inferred gas volume, at 25 °C/atm.			
In gas hydrates	1,000 TCF (3 x $10^{13} m^3$)	200 TCF (6 x $10^{12} m^3$)	1,200 TCF (4 x $10^{13} m^3$)
In trapped free gas	N/A	N/A	100 TCF (3 x $10^{12} m^3$)

INTRODUCTION

The Gulf of Mexico, often categorized as a small ocean basin (Menard, 1967) or a Mediterranean type sea (Garrison and Martin, 1973) covers an area of more than 1.5 million km^2. It has many of the geomorphologic features but few of the geologic features of a large ocean. The geologic evolution of the Gulf of Mexico is very distinctive and is dominated by salt deposition and salt diapirism at an unprecedented scale.

The scope of this report is limited, on the basis of documented and inferred gas hydrate occurrences, to the western half of the Gulf of Mexico, with emphasis given to the geology of the northwestern and western continental margins (the Texas - Louisiana and east Mexico continental slopes) and the central, deep Gulf of Mexico Basin (the Sigsbee Plain and Bay of Campeche).

The western Gulf of Mexico can be readily subdivided into the following physiographic regions (Figure 1).

- the central Gulf of Mexico basin: Sigsbee Abyssal Plain, Sigsbee Knolls, distal portions of the Mississippi Deep Sea Fan.

- the northwestern margin: Mississippi Delta and Mississippi Deep Sea Fan, Texas - Louisiana Coastal Plain, Continental Shelf and Continental Slope and the Sigsbee Escarpment,

- the western margin: Rio Grande Continental Slope and Perdido Escarpment, East Mexico Continental Shelf (Tamaulipas and Veracruz) and Continental Slope (Mexican Ridges), Veracruz Tongue, Campeche Knolls (Campeche Continental Slope), Yucatan Continental Shelf (Banco de Campeche) and its frontal scarp, the Campeche Escarpment.

Gas hydrates in oceanic sediments have been reported or inferred from various regions of the Gulf of Mexico. The first indication of the presence of gas hydrates was provided in 1970 by the Deep Sea Drilling Project, DSDP, Leg 10, when unusual degassing features were observed from gas-rich cores recovered from sediments collected from the deep-water Sigsbee Plain and Gulf of Campeche. At the time of core recovery, these observations were perplexing, but it was realized in retrospect that the core had intersected, in all likelihood, some gas hydrate-bearing intervals (Worzel and Bryant, 1973).

Seismic bottom simulating reflectors (BSRs) were interpreted as possible gas hydrate reflectors by Buffler et al. in 1979 from the Mexican Ridges anticlinal ridges, located along the continental slope of the western Gulf of Mexico. This interpretation was apparently confirmed in some detail by Hedberg (1980) who specifically discussed BSRs on unlocated seismic profiles BSRs as gas hydrate reflectors.

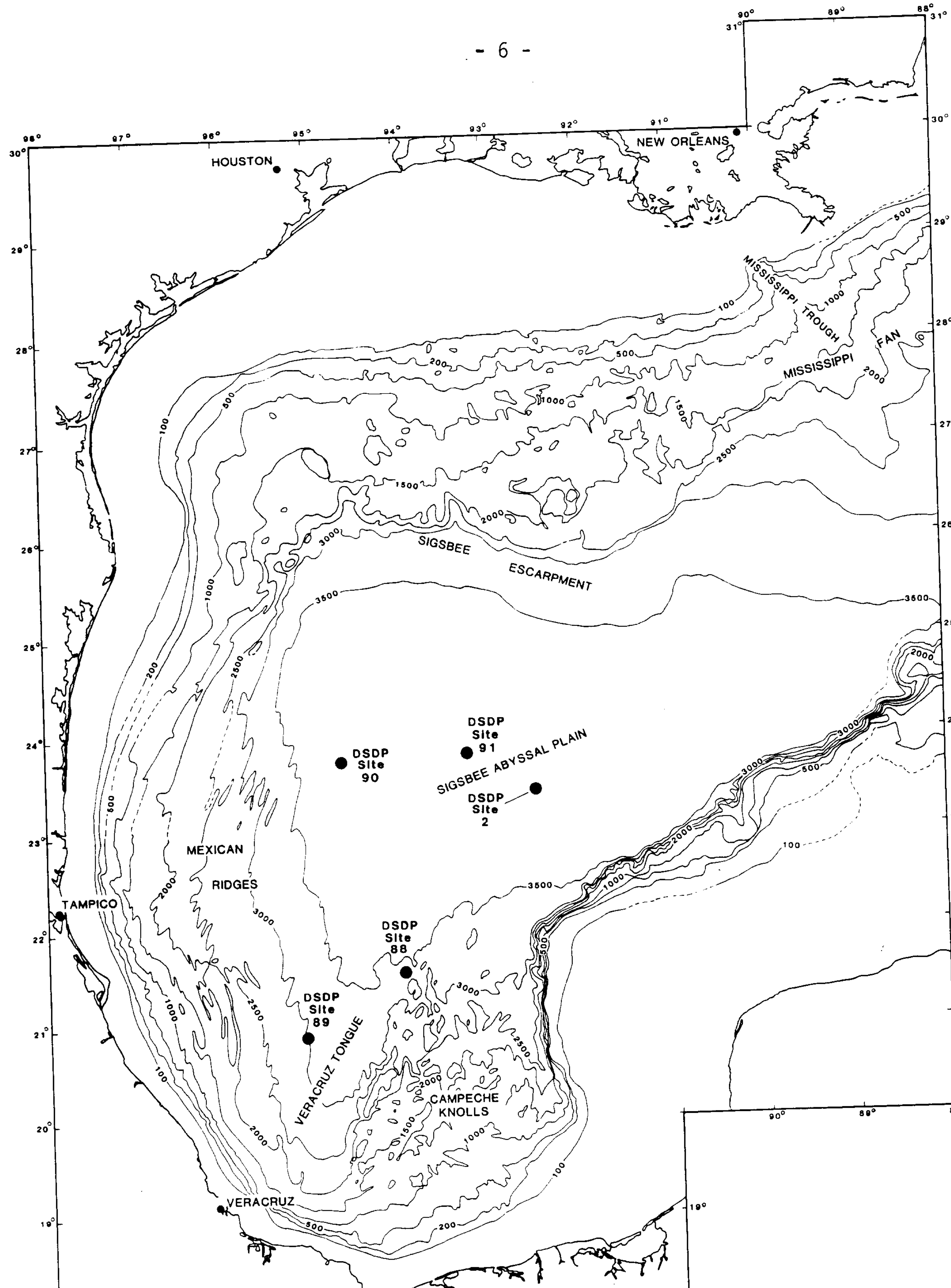

Figure 1. **BATHYMETRIC MAP OF THE WESTERN GULF OF MEXICO**

Modified after Bryant et al., in Buffler et al., (1984)

These seismically inferred gas hydrate occurrences from the western Gulf of Mexico were the basis for the U.S. Department of Energy - Morgantown Energy Technology Center's (DOE - METC), designation of the Gulf of Mexico as an oceanic sediment gas hydrate region (specifically DOE - METC sites 17 and 18).

Since 1983 gas hydrates have been recovered during coring operations from the Louisiana Continental Slope, by a Texas A & M University party from the Green Canyon, Garden Banks, and Mississippi Trough lease areas. In late 1983, gas hydrates were also encountered and recovered by the DSDP leg 96 scientific party from the Orca Basin, an intraslope basin of the Louisiana lower continental slope, bringing the number of gas hydrate occurrences of the Louisiana continental slope to eight (Brooks and Bryant, 1985a).

Acknowledgements

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Personnel from the National Geophysical Data Center, Boulder, Colorado, were very helpful in locating seismic and geochemical data on the Gulf of Mexico. William R. Bryant provided us with literature and seismic profiles of the region. Margaret Krason drafted the figures for this report.

PART I

BASIN ANALYSIS

The geologic evolution of the western Gulf of Mexico and its margins is distinctive and presents many unresolved and controversial problems. Briefly, the salient points and problems pertaining to the evolution of the Gulf of Mexico are:

1. The deep crustal nature and early geologic evolution of the central Gulf of Mexico Basin is not thoroughly understood. The recent concept of rifted continental crust rimming a central deep basin floored by oceanic crust has proven to be a useful model. The complete lack of evidence for a spreading center in conjunction with extremely thick sedimentary accumulation since Jurassic time, indicate that the Gulf of Mexico Basin, even if floored with oceanic crust, departs significantly in its evolution from a typical rifted oceanic basin.

2. The role of salt diapirism at a scale apparently unprecedented in the geologic record is the major and most controversial feature of the western Gulf of Mexico and represents the connecting links between its various geographical components (Figure 2).

   The geologic setting and the mode and age of deposition of the evaporitic unit or units, are crucial points in any hypotheses on the geologic evolution of the western Gulf of Mexico and its margins. Recent models contend that the evaporitic sequences were deposited within separate but partly connected extensional basins during Middle to Late Jurassic time over red bed continental sequences in graben-type structure. These red beds and evaporitic sequences are thought to represent a syn-rift rock assemblage deposited over passive continental margins in the process of deep seated attenuation and extension at a higher crustal level.

   Salt diapirism, documented from the northwestern (Texas - Louisiana Continental Slope and inland areas) and southern (Campeche Knolls and Isthmus saline basin) occurs on a giant scale (Figure 2). It has been hypothesized in the case of the northwestern margin that the thick salt formation was mobilized by the increase of lithostatic pressure due to the rapid seaward progradation of a thick clastic apron during Tertiary and Quaternary time. The massive Sigsbee Escarpment represents a huge salt bulge built up by lateral flow of the salt at a regional scale. A similar allochthonous interpretation

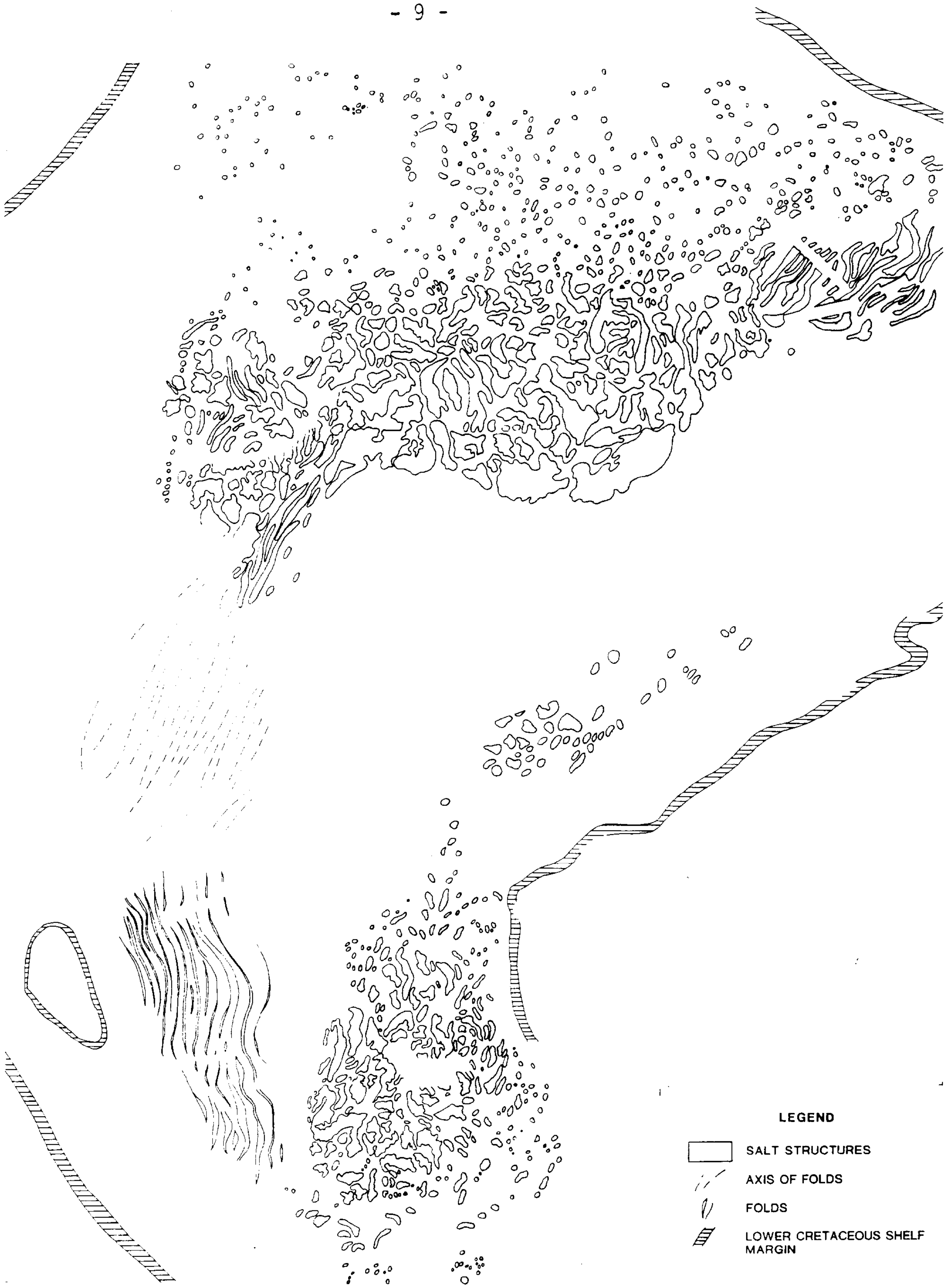


Figure 2. TECTONIC FEATURES OF THE WESTERN GULF OF MEXICO

Modified after Locker and Sahagian, in Buffler et al., (1984)

has been convincingly applied to the frontal edge of the diapiric Campeche Knolls province.

3. The nature of the escarpments which border the northern and southern margins of the Gulf of Mexico and dominate the deep Gulf of Mexico Basin (Sigsbee Abyssal Plain) is comparatively well understood (Figure 1). These geomorphologic escarpments are of two different kinds, each representing unusual and spectacular geologic features. The Sigsbee and Perdido Escarpments of the northwestern margin represent the physiographic expression of giant diapiric salt walls which are located at the seaward edge of a salt diapir-dominated geologic province (Texas - Louisiana Continental Slope). The abrupt Campeche Escarpment of the southern margin of the Gulf of Mexico represents the constructional edge of the Cretaceous carbonate Yucatan Peninsula.

4. The early history of the Gulf of Mexico could only be comprehended within a conceptual framework previously developed from other regions. The first clear picture emerges for the middle Cretaceous when nearly the entire Gulf of Mexico was rimmed by one of the largest carbonate platform complexes documented from the geologic record (Figure 2). The constructional edge of the platform along the northern and western margins were bypassed and buried later but prevailed up to the present as the southern margin (Yucatan - Campeche carbonate platform). The fringing, near sea level, Cretaceous carbonate platform complex and its inland evaporitic environment nearly enclosed the deep Gulf of Mexico Basin which was the site of presumably calcareous, hemipelagic sedimentation. The Late Cretaceous heralded the final separation of the Gulf of Mexico into a central, deep-water basin rimmed by shallower carbonate, then clastic margins, which persisted up to the present.

5. The delineation and size reduction by seaward encroachment of the margins of the deep Gulf of Mexico Basin in mid-Cretaceous time was accentuated during the Late Cretaceous through Quaternary by clastic shelf edge progradation of the northern and western margins when a drastic change in sedimentation type from shallow marine carbonate to deltaic clastic took place. The phenomenon is strikingly displayed in the Texas margin where a 400 km deltaic progradation is recorded since the drowning and burial of the middle Cretaceous shelf-edge carbonate platform. Literally speaking the central Gulf of Mexico Basin has been steadily shrinking since. The deep central basin was not a starved basin, but received distal turbidites and hemipelagites in excess of 7,000 to 8,000 m in thickness up to the present. The steady subsidence pattern implied by this type of sediment thickness is at variance with the subsidence evolution for an oceanic basin.

6. The Quaternary witnessed an accelerated clastic margin progradation linked to periods of sea level lowstands when increased clastic supplies were supplied directly to the shelf margins. The Mississippi River breached the northern Louisiana shelf; this event

represents the first documented encroachment of a major clastic system within the deep Gulf of Mexico Basin. The Mississippi River built a deep sea fan as much as 3,000 m thick in its median constructional part whose distal fringe abutted the toe of the Campeche Escarpment.

The Central Gulf of Mexico Basin

Origin

In a critical review of hypotheses on the evolution of the Gulf of Mexico Basin, Uchupi (1975) divided previous investigators in two schools of thought: the staticist school and the mobilist school.

The concept of permanence of the Gulf of Mexico Basin was last advocated by Meyerhoff and others in the early 1970s (Meyerhoff, 1967; Paine and Meyerhoff, 1970; Meyerhoff and Meyerhoff, 1972). They regarded the oceanic crust-floored deep basin as a subsided plate, at least as old as Late Mississippian to Early Pennsylvanian. Meyerhoff's investigations took into account the geology of the borderlands and was supported by a Potassium-Argon radiometric date from a purple siltstone dredged from a knoll of the Sigsbee group, which yielded a Pennsylvanian age (Pequegnat et al., 1971). This age has been doubted or disregarded by most subsequent investigators of the plate tectonics mobilist school (Watkins et al., 1975, for example).

The mobilist school has enjoyed widespread popularity with the advent of plate tectonics theory. All hypotheses seem to involve some sort of crustal rifting period in the late Triassic to Jurassic and/or major translation and rotation of crustal blocks along 'megashears', wrench faults, or transcurrent faults. The central part of the Gulf of Mexico Basin, floored by deep oceanic crust is related to the opening of the Atlantic Ocean or as a translated fragment of an older Pacific Ocean (Yarborough, 1968). The concept of a Caribbean microplate resulting from the decoupling of the two major North and South American plates along major transcurrent structures is still speculative (Uchupi, 1975). The articles and abstracts on the 1980 symposium on the origin of the Gulf of Mexico (Pilger, 1980) present a recent overview of current theories.

Since the pioneering seismic refraction works of Ewing et al. (1955, 1960, 1963), Cram (1960), and Antoine and Ewing (1963), the oceanic crustal nature of the basement of the deep central part of the Gulf of Mexico Basin has been established. More recent seismic refraction work has supported this interpretation (Hales et al., 1970; Dorman et al., 1972; Buffler et al., 1981; Ibrahim et al., 1981; Ibrahim and Uchupi, 1983). One of the great difficulties encountered in trying to interpret the deep crustal layer of the Gulf of Mexico Basin in terms of plate tectonics theory is the absence of any well defined, linear, symmetrical, magnetic anomalies, which could be interpreted as a spreading center within the central Gulf of Mexico (Martin and Case, 1975; Buffler et al., 1981). The oceanic type basement is inferred exclusively from seismic refraction velocity studies. This seismically defined oceanic type crustal layer is overlain by an extremely thick sedimentary cover averaging 10,000 m thick.

Syn-Rift Sequence

The first definition of seismic units through the deep Gulf of Mexico Basin was provided by Ladd et al. (1976). They recognized faint, noncontinuous, tilted reflectors under a well defined gulfwide seismic unit. This latter unit termed the Challenger unit was inferred to represent the salt-bearing sequence of the deep Gulf basin and was assigned a Jurassic age. Similar terminology was adopted by Watkins et al. (1975, 1976a, 1976b, 1978) and Worzel and Burk (1979), among others. A modification of this seismic nomenclature by Shaub et al. (1984) is used in this report (Table 2).

The pre-Challenger tilted sequence was inferred to represent a pre- to Early Jurassic clastic syn-rift assemblage. A Paleozoic age for some of these rocks was recognized as possible in view of the previously mentioned Pennsylvanian rock dredged from the Gulf. However, most investigators have regarded the dating as spurious and have accepted a Late Triassic to Early Jurassic age for these strata.

Modern seismic reflection profiling through the deep Gulf of Mexico Basin led Buffler et al. (1980, 1981) to a new synthesis, introducing the concept of thinned, rifted continental crust (Figures 3 - 5). In this interpretation, the former view of syn-rift, continental clastics and volcanics deposited into extensional, graben structures was expanded, based on deep seismic events identified north and northeast of the Campeche Escarpment. Tilted continental basement blocks, suggesting rotation along listric faults, and thick syn-rift sediment fills were inferred using the extensional crustal model documented by Montadert et al. (1979) from the Bay of Biscay for the Atlantic edge of the European craton. A similar model was applied to the northern margin of the Gulf of Mexico where extensive salt diapirism along the continental slope has completely obscured pre-salt unit configuration.

The zone of contact between oceanic and continental crust has been a subject of much debate. The concept of wide rifted, attenuated continental crust (transitional crust of recent authors) has helped clarify this subject. Still disagreements and different interpretations persist as a recent review by Hall (1983) demonstrates. This report conforms with the synthesis presented by Buffler et al. (1980, 1981) and the generalization that the present seaward outline of the diapiric salt provinces of the southern and northern margin reflect the outer extent of the rifted continental crust assemblage (Figures 6 - 7).

Salt Sequence

Historical Review. Salt diapirs or salt and gypsum cored anticlines have been identified on the mainland, from Mississippi in the United States, to the Isthmus of Tehuantepec (Veracruz and Tabasco in Mexico, and farther east, in Guatemala). In the 1960s, the extent of the salt diapir fields was recognized as particularly extensive and of almost circum-Gulf nature (Murray, 1961, 1966; Halbouty, 1967). The extent of salt diapirism in offshore Texas and Louisiana was recognized, but it was not until 1969 (Lehner, 1969) that the final proof of the salt extent was established after a number of U.S. continental slope domal features were drilled and confirmed to be cored by salt (Figure 2).

TABLE 2.

DEEP GULF OF MEXICO SEISMIC UNITS
After Buffler et al. (1984)

Unit	Age	Suggested Depositional Environment Depocenter/Source
Sigsbee	Pleistocene	Abyssal submarine fan and other northern-source mass-transport deposits in eastern 2/3 of Gulf, contributed by Pleistocene Mississippi River, mostly suspension deposits in west; some fine-grained turbidites also derived from Mexican rivers in western basin.
Cinco de Mayo	Late Miocene to Pliocene	Abyssal terrigenous and biogenic ooze. No major depocenter in deep Gulf; sediments thicken slightly in northern and southwestern Gulf. Most of clastic supply may be trapped by sedimentary deformation along northern and western margins.
Upper Mexican Ridges	Middle Tertiary(?) to Late Miocene	Predominantly deep marine distal sands, silts, muds. Western margin progradation and sedimentation continues from western margin. A northeastern depocenter is attributed to the ancestral Mississippi River.
Lower Mexican Ridges	Early(?) to Middle(?) Tertiary	Predominantly deep marine distal clastic sediments. Broad western depocenter attributable to ancestral Texas and Mexican rivers. Considered a continuation and progradation of the sedimentation pattern of the underlying Campeche unit. Clastics are distributed throughout the entire deep basin.
Campeche	Upper Cretaceous to Early Tertiary(?)	Predominantly deep marine distal clastics in west pelagic in east. Western depocenter source is probably Rio Grande Embayment.
Challenger	Middle Jurassic(?) to Middle Crataceous	Unit immediately overlies acoustic basement (oceanic and transitional crust). Predominantly deep marine sediments in central Gulf. Evaporites, shallow then deep marine along Campeche Escarpment. Central Gulf depocenter has apparent source in Campeche region.

FIGURE 3. Generalized geological cross-section through the Gulf of Mexico from the Texas - Louisiana coastal plain to the Yucatan Peninsula, is located in the pocket at the end of the report.

FIGURE 4, Generalized geologic cross-section through the Gulf of Mexico from eastern Mexico through the Mississippi Fan, is located in the pocket at the end of the report.

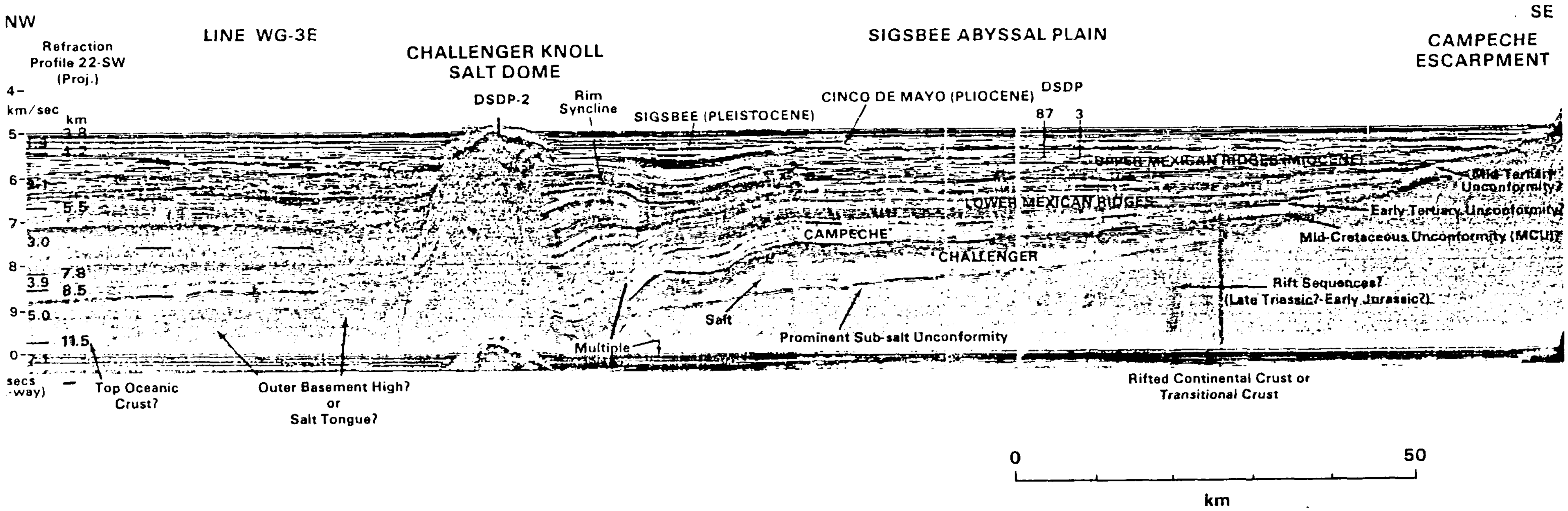


Figure 5. SEISMIC PROFILE THROUGH THE DEEP-CENTRAL GULF OF MEXICO

From Buffler, in Bally (1983)

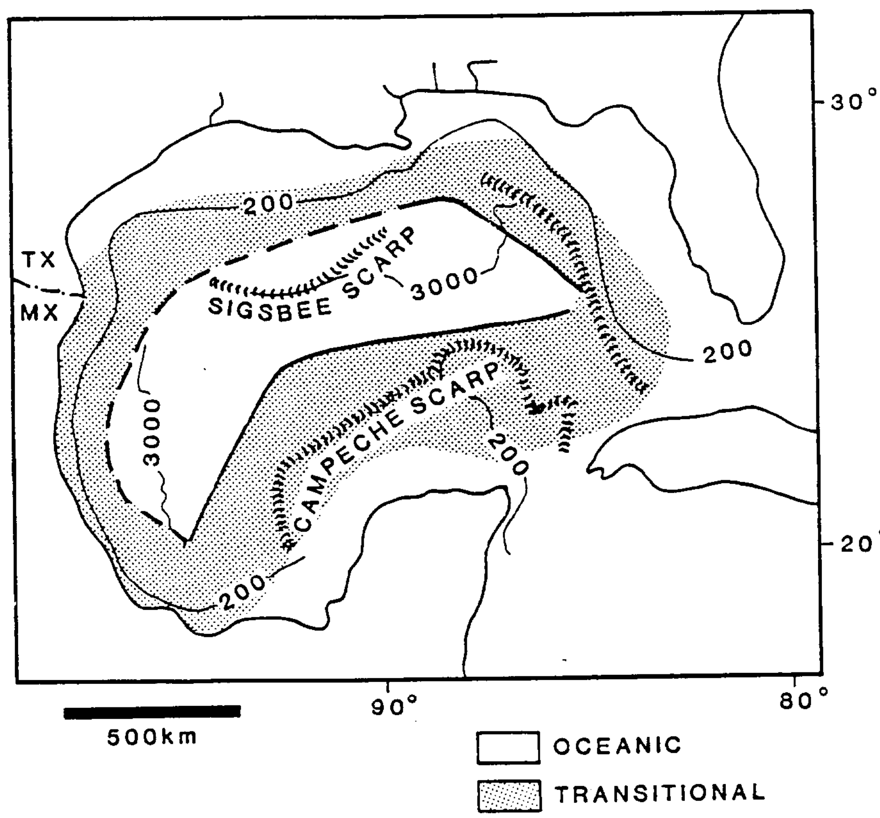


Figure 6. INFERRED DISTRIBUTION OF OCEANIC VERSUS RIFTED TRANSITIONAL CRUST IN THE CENTRAL GULF OF MEXICO

After Buffler et al., 1980

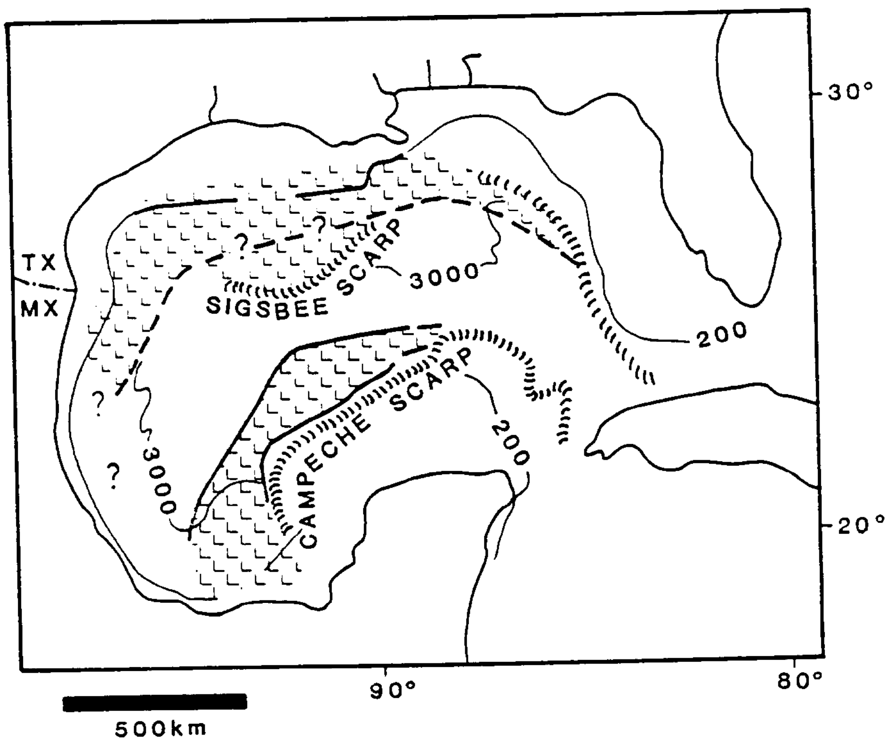


Figure 7. INFERRED EXTENT OF THE SALT AROUND THE DEEP CENTRAL GULF OF MEXICO

After Buffler, 1980

In the same year, the Deep Sea Drilling Project, Leg 1 (1968) confirmed the suspicion, first voiced by Ewing et al. (1962) and Talwani and Ewing (1966), that the Sigsbee Knolls of the south central deep Gulf of Mexico were also salt diapirs, by encountering a salt cap rock atop the Challenger Knoll (Ewing et al., 1969a, 1969c; Burk et al., 1969; Figure 5). The continental slope of the Bay of Campeche was also recognized as a giant salt diapiric province mirroring the geologic setting of the Texas and Louisiana continental slope and corresponding to the offshore extension of the Isthmus salt province of central Mexico (Worzel et al., 1968; Ballard and Feden, 1970; Ensminger and Matthews, 1972; Figure 2). Syntheses papers by Antoine and Bryant (1969), Antoine and Pyle (1970), and Antoine (1972) inferred an offshore salt belt rimming the entire southern, western and northern margin of the western Gulf of Mexico, inferring the presence of thick salt under the eastern Mexican continental shelf and continental slope following the interpretation of Bryant et al. (1968).

The geologic study of the deep Gulf basin and the extent of the salt culminated in the mid to late 1970s with a series of articles on seismic stratigraphy by Ladd et al. (1976), Watkins et al. (1975, 1976a, 1977, 1978) and Worzel and Burk (1979) which have been refined by Powers (1981), Foote et al. (1983), Shaub et al. (1984), and Buffler et al. (1984).

The salt-bearing seismic unit, named the Challenger from Challenger Knoll, was the first recognized widespread gulfwide stratigraphic unit. Its top is deformed by some of the Sigsbee Knolls, and some seismic sections, according to Worzel and Burk (1979) show some salt diapirs originating from the basal portion of the Challenger. The Challenger unit onlaps southward onto the attenuated continental basement of the Campeche Escarpment (Figure 5), but its extent has been debated and the amount of massive salt within the seismic unit has also been questioned. It is still debatable whether a thick salt unit covers part or all of the deep Gulf basin. The Challenger is shown in some seismic sections as not reaching the Sigsbee Escarpment, an allochthonous salt ridge complex encroaching into the deep Gulf. Conversely, the extent of the Challenger and the salt near the east Mexican slope and the Mexican Ridges is uncertain.

Age. The age of the salt unit from the deep Gulf of Mexico Basin, the U.S. continental slope and from the Gulf of Campeche has been assigned, since the time of its discovery, as Jurassic.

The evidence regarding the age of the salt formation is mostly derived from circum-Gulf offshore and inland salt basins. The only direct evidence from the deep Gulf of Mexico Basin derived from palynomorphs study collected from the cap rock of Challenger Knoll drilled as Site 2 of DSDP Leg 1 (Ewing et al., 1969c) which yielded a probable Late Jurassic age (Burk et al., 1969; Kirkland and Gerhard, 1971; Cusminer, 1978) in accordance with the data gathered from the circum-Gulf salt basins (Figure 5). Early investigators (Ewing et al., 1969; Burk et al., 1969; and others) thereby deduced that the salt of the deep basin should broadly correlate with the Louann salt of the inland areas of the northwestern margin. Similarly, the salt coring the diapirs of the Gulf of Campeche and the Sigsbee Knolls was correlated with the onland diapiric field of central Mexico, the Isthmus saline basin.

The Louann salt was considered to be Early to Middle Jurassic in age and is presently thought to be late Callovian to late Oxfordian on the basis of local stratigraphic correlation and on the basis of long-range correlation with the evaporitic formation known from the Coahuila and Nuevo Leon states of northeastern Mexico (Imlay, 1980).

The Minas Viejas Formation, which is documented to rest in some places above Triassic and Early to Middle Jurassic red bed sequences (La Joya Formation and equivalents) was identified in the Parras and Sabinas basins and in the anticlinal cores of some large decollement folds of the Coahuila Marginal fold belt. It is composed of a lower massive salt unit overlain by massive anhydrite and is 500 to 1,000 m thick. It is stratigraphically overlain by the Zualaga Limestone, a lithostratigraphic equivalent from the Smackover Formation of the U.S. inland Gulf Coast (Imlay, 1943; Wall et al., 1961; Ewing and Antoine, 1966; Murray, 1966; Weidie and Murray, 1967; Halbouty, 1967 and 1979; Weidie and Martinez, 1970; Kirkland and Gerhard, 1971). According to Imlay (1980), the Minas Viejas Formation is late Callovian to early Oxfordian.

The salt of the diapiric continental slope of the Gulf of Campeche is obviously related to the onland salt basin and oil-producing area of Central Mexico known as the Isthmian saline basin (Acevedo, 1980; Acevedo and Dautt, 1980). This vast salt-bearing region was referred to as the Great Campeche Salt Basin by Viniegra (1981).

The determination of the age of the salt formation has been a vexing problem (Imlay, 1943; Halbouty, 1967 and 1979; Contreras and Castillon, 1968; Kirkland and Gerhard, 1971; Viniegra, 1971 and 1981; Bishop, 1980; among others). Schematically, a major evaporitic sequence is documented from the central part of the Isthmian saline basin, consisting of two thick, evaporitic salt-anhydrite cycles separated by a red clastic interval of the Todos Santos Group. Depositional thicknesses are in the range of many hundreds to thousands of meters. The Todos Santos Group encompasses a thick red bed sequence widespread throughout Central America. It is obviously diachronous (time-transgressive) at a regional scale, spanning the Early to Late Jurassic and is as young as Early Cretaceous in Guatemala. It appears to have been deposited, at least in part, in elongate fault controlled troughs. It rests along the exposed southern margin of the basin on a composite deformed basement assemblage and locally interfingers with volcanics which have yielded a Late Jurassic radiometric date. This widespread continental clastic complex interfingers northward at a regional and local scale with basinal evaporites. Only two of these evaporitic intervals appear to have regional extent. The salt-anhydrite units interfinger with and are overlain by more basinal, occasionally open marine limestone facies (Chinameca Limestone and lithostratigraphic equivalents) which permits paleontological dating. All the evidence at hand indicates that the red bed, evaporite sequence is pre-Kimmeridgian. In the Reforma oil field area, from which subsurface data are plentiful, the main evaporite unit is more precisely dated at its top as pre-late Oxfordian. In the offshore producing area of the southern Gulf of Campeche, a slightly older age is documented (Callovian). The presence of a younger but less extensive evaporitic cycle is also inferred from the offshore area. These Middle to Late Jurassic ages are supported by palynomorph studies (Cousminer, 1978).

A Late Jurassic age appears to be well established and accepted by all investigators for the main salt-bearing unit from the central and circum-Gulf of Mexico. But, significant Late Jurassic and Early Cretaceous evaporites, mostly anhydrite and gypsum, are known from various inland regions bordering the Gulf of Mexico and may possibly have offshore equivalents. Cretaceous anhydrite and gypsum formations were identified in the back reef, restricted marine facies from the Cretaceous carbonate platforms from western Mexico (Golden Lane, Valles - San Luis Potosi, and Coahuila platforms) and are locally involved in decollement tectonics. The larger Early Cretaceous circum-Gulf evaporitic basin was located in Chiapas and northern Guatemala, partly overlapping over the former clastic-evaporitic Isthmus saline basin (Viniegra, 1971 and 1981; Bishop 1980). This restricted marine anhydrite and dolomite shallow water lagoonal series is thick; one well intersected 3,000 m of evaporites.

Similar restricted marine, lagoonal anhydrite and dolomite units are documented in the carbonate sequence of the Yucatan Peninsula. These evaporitic cycles are not of regional significance but have been reported from the Cretaceous and the Paleogene. All investigators agree that these anhydrite units have not been involved in postdepositional diapirism and that they do not have lithologic equivalents into the deep Gulf basin. Also, similar restricted marine anhydrite units of widespread extent are known from the Late Jurassic to Early Cretaceous formations of the Mississippi Embayment.

Depositional Environment The Jurassic salt deposition episode is considered to have occured in extensional, separate but intermittently connected basins developed on a continental crust assemblage in the process of deep-seated attenuation and high level extensional rifting (Buffler et al., 1980 and 1981). The salt deposition marked the near end of this period of rifting and extensional tectonics. It was followed in the Late Jurassic or Early Cretaceous by a period of rifting and generation of new oceanic crust presently flooring the very central part of the Gulf of Mexico. This newly created crust passively subsided along with the now separated, attenuated continental crust of the northern and southern margins. However, Watkins et al. (1976 and 1978) envisaged all the Gulf of Mexico floored by oceanic crust on which, after a period of clastic deposition, a thick blanket of salt was uniformly deposited throughout the deep basin.

Post-Rift Sequence

A minor rifting phase followed the deposition of marine salt into individual subsiding basins of the Late Jurassic (Buffler et al., 1980 and 1981). A major rift, floored by newly generated oceanic crust, opened and the continental margins separated. This event apparently was fairly brief as evidenced by crustal separation of only several hundred kilometers. The separated salt basins accounted for the present symmetrical configuration of the northern and southern margins.

After sea floor spreading in the central Gulf of Mexico abated, the basin and its margins subsided as the lithosphere cooled and contracted. Early deformation of salt and overlying sediments by gravity flowage was probably triggered by basin subsidence. This was documented in the southern Campeche margin by salt pillow structures and associated folded sediments overlying an undeformed middle Cretaceous regional unconformity (Buffler et al., 1980; Figures 3 - 5). Differential thermal subsidence patterns led to a differentiation of a central deep Gulf basin, rimmed by shallow water platforms. An Early Cretaceous shallow water carbonate bank edge, controlled in part by a tectonic hinge zone separating continental blocks from the thinner transitional crust, developed on the southern margin.

By early Late Cretaceous time, the carbonate bank of the southern Campeche margin had developed into a high constructional escarpment rimming the deep-water basin where fine-grained hemipelagic carbonate sediments were deposited (Campeche seismic unit; Figure 5). The seaward edge of the carbonate platforms extended continuously all around the Gulf of Mexico Basin (Figure 2). The northern and northwestern carbonate shelf edge margins were not as developed as their southern counterparts and were overlapped and buried by a major increase in clastic sediment input following orogeny in the western North American continent starting in the Late Cretaceous.

This change in sedimentary patterns led to a progradation of deltaic complexes and associated turbidites and slumps into the open Gulf of Mexico. The central Gulf of Mexico Basin steadily subsided and was filled by a thick sequence of fine-grained mudstones and foraminiferal ooze. This clastic sequence is 3,000 m thick in front of the Campeche Escarpment and 9,000 to 10,000 m thick at the edge of the Sigsbee Escarpment and the southern Mexican Ridges. This sequence has been subdivided into widespread units based on characteristic seismic reflections (Watkins et al, 1975, 1976a, 1976b, 1977, 1978; Ladd et al., 1976; Worzel and Burk, 1979; Power, 1981; Foote et al., 1983; Shaub et al., 1984; Buffler et al., 1984; Figures 8 - 13). The isopach map of the total sequence (Figure 8) clearly shows the increase in sediment thickness from the base of the Campeche Escarpment to the Sigsbee Escarpment to the north and the Mexican Ridges to the west. The thinning by overlap of all except the most recent units (late Miocene to Holocene) over the toe of the Campeche Escarpment is very well displayed in many seismic profiles (Figure 5). The steady increase in sediment thickness toward the northwest reflects a regional tilt of the deep Gulf Basin the the Tertiary, and is still reflected in the present sea floor gradient of the Sigsbee Plain. Isopach mapping of individual units documents the main source of sediments as being from the west and northwest margins (Shaub et al, 1984; Buffler et al., 1984; Figures 9 - 13).

By the late Miocene, although the influx of sediment from the southwestern portion of the Gulf was still very important, the influence of the ancestral Mississippi River system became significant; the main depocenter shifted to southern Louisiana. This deposition pattern climaxed during the Pleistocene with the building of the deep sea Mississippi Fan (Figures 12 - 13).

The upper part of the clastic infill of the Sigsbee Abyssal Plain has been tested by the Deep Sea Drilling Project, Leg 1 and 10 (Ewing et al., 1969a; Worzel et al., 1973a). The deepest horizons drilled were from the mid-Miocene; detailed supporting shallow seismic profiling shows local overlaps and pinch outs of lower units. Laminated turbidites, silty laminated or blocky mudstone, and burrowed nannofossil-bearing hemipelagic mudstones have been recovered.

Worzel and Bryant (1973) noted that each of the three sites drilled during DSDP, Leg 10 into the abyssal plain of the central and southern central Gulf of Mexico encountered gassy upper sections and unusual core degassing features characteristic of dissociating gas hydrates in Miocene to Pleistocene sections.

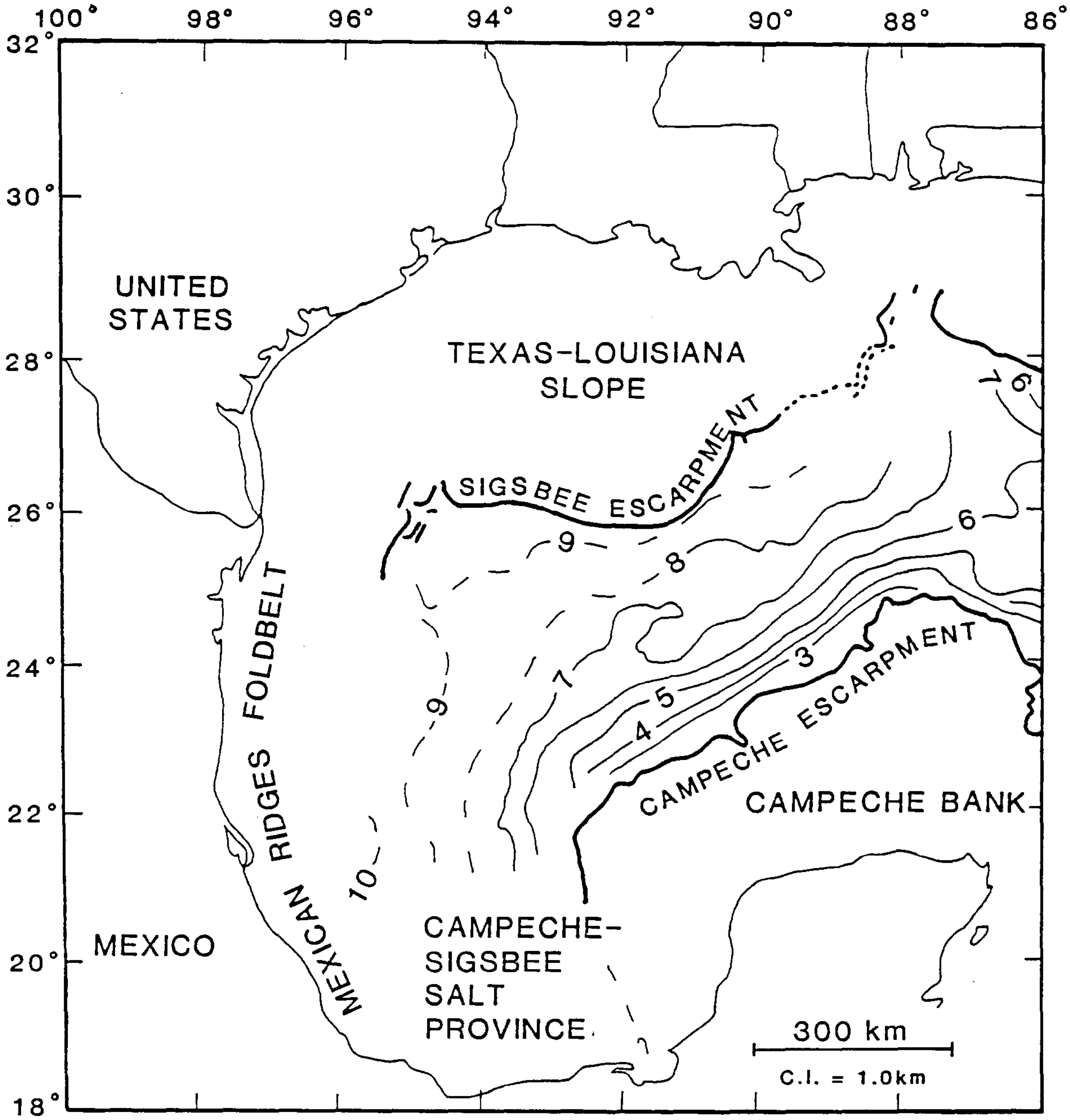


Figure 8. **TOTAL SEDIMENT ISOPACH MAP OF DEEP GULF OF MEXICO BASIN**

After Shaub et al. (1984)

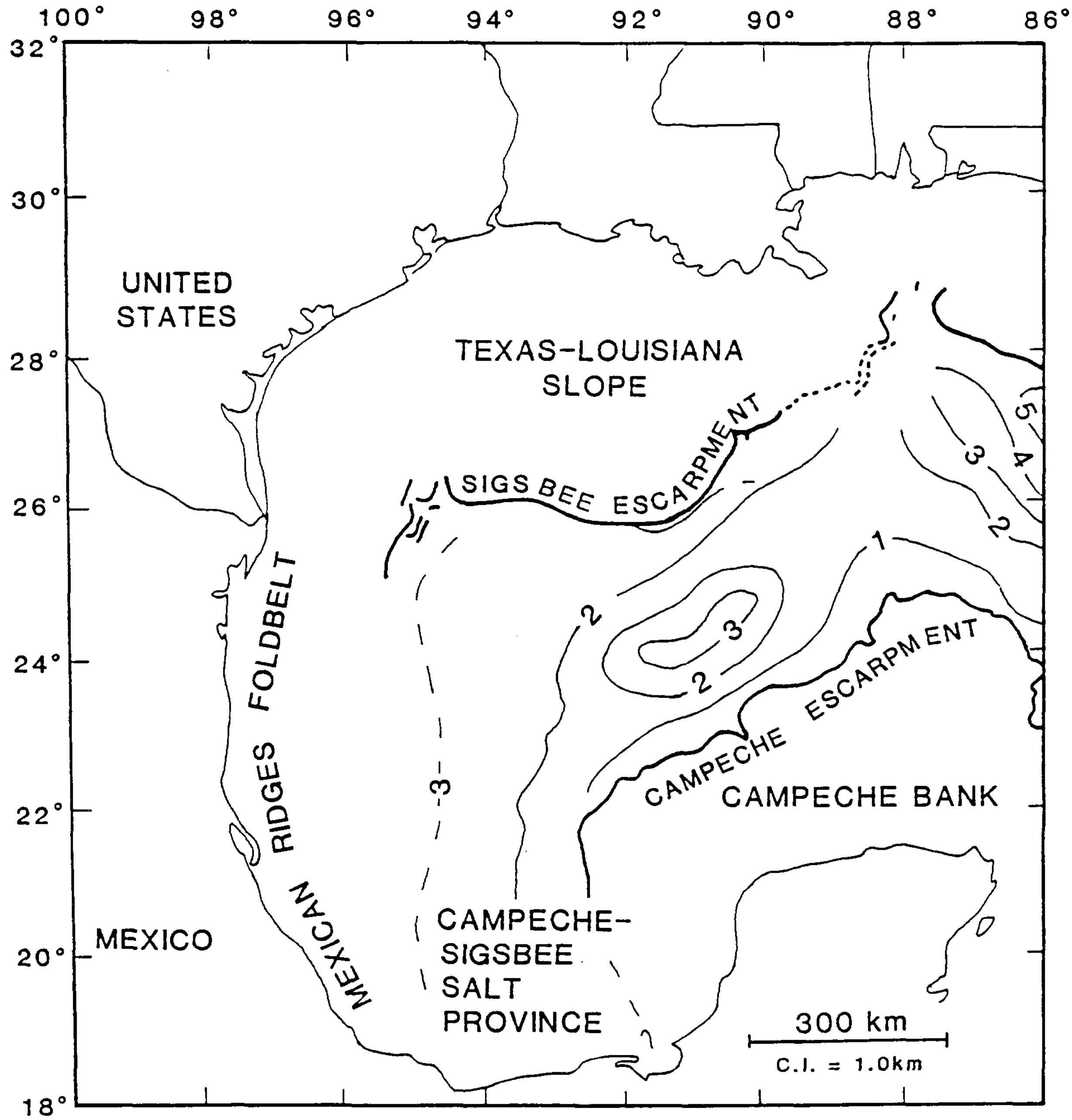


Figure 9. GENERALIZED ISOPACH MAP OF JURASSIC (?) TO MIDDLE CRETACEOUS CHALLENGER UNIT

After Shaub et al. (1984)

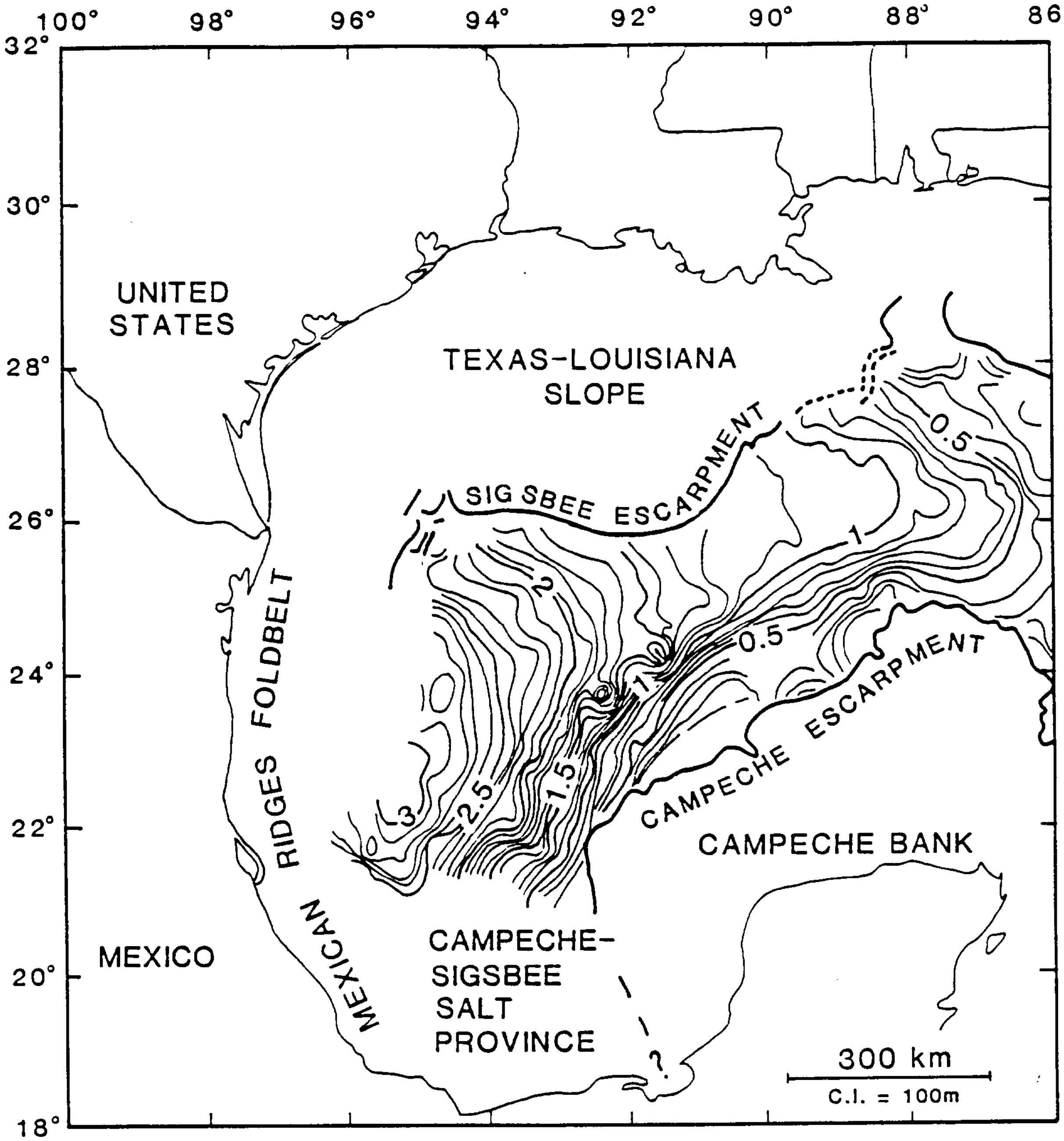


Figure 10. ISOPACH MAP OF MIDDLE CRETACEOUS TO EARLY TERTIARY (?) CAMPECHE UNIT

After Shaub et al. (1984)

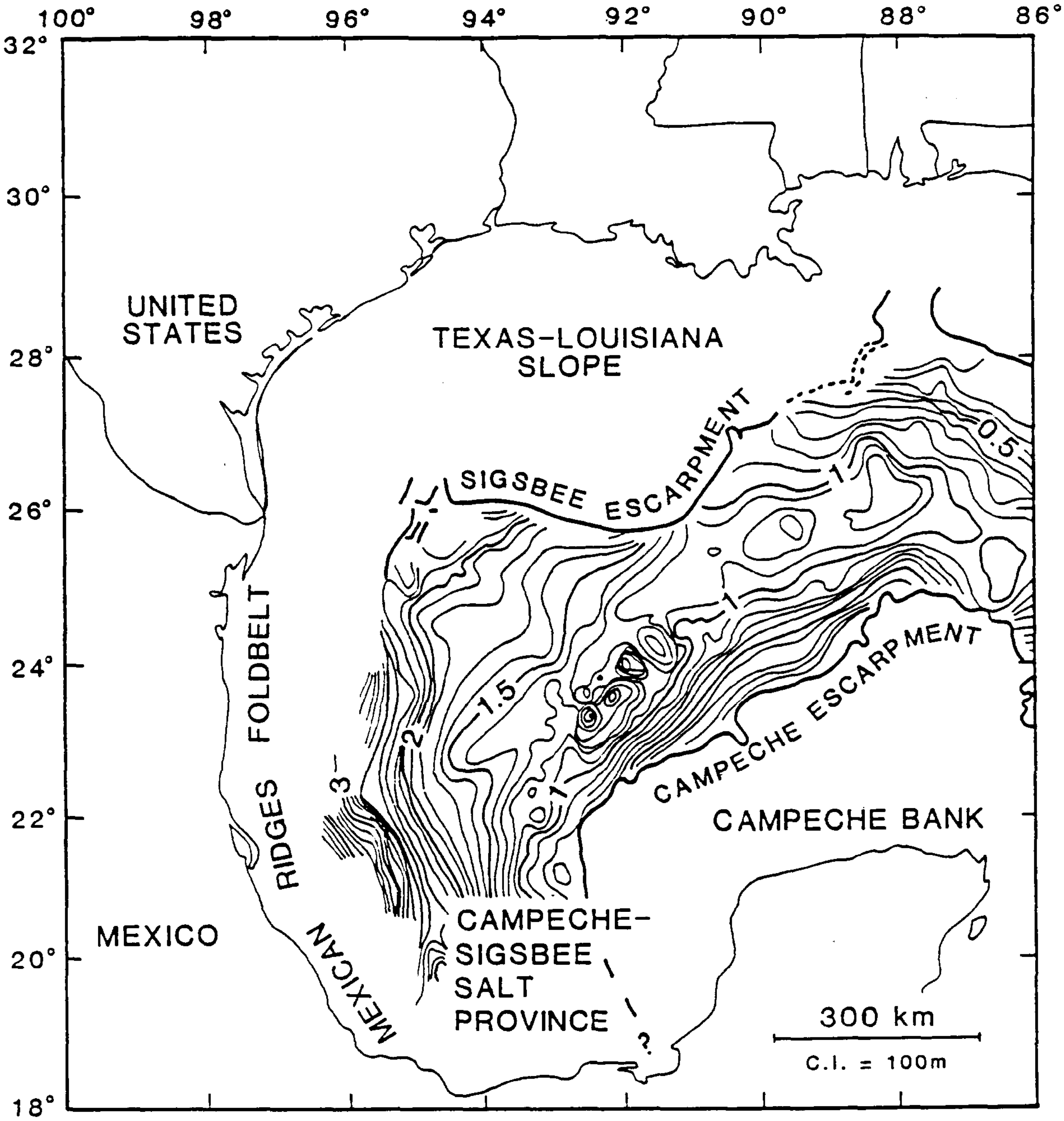


Figure 11. ISOPACH MAP OF EARLY (?) TO MIDDLE TERTIARY (?) LOWER MEXICAN RIDGES UNIT

After Shaub et al. (1984)

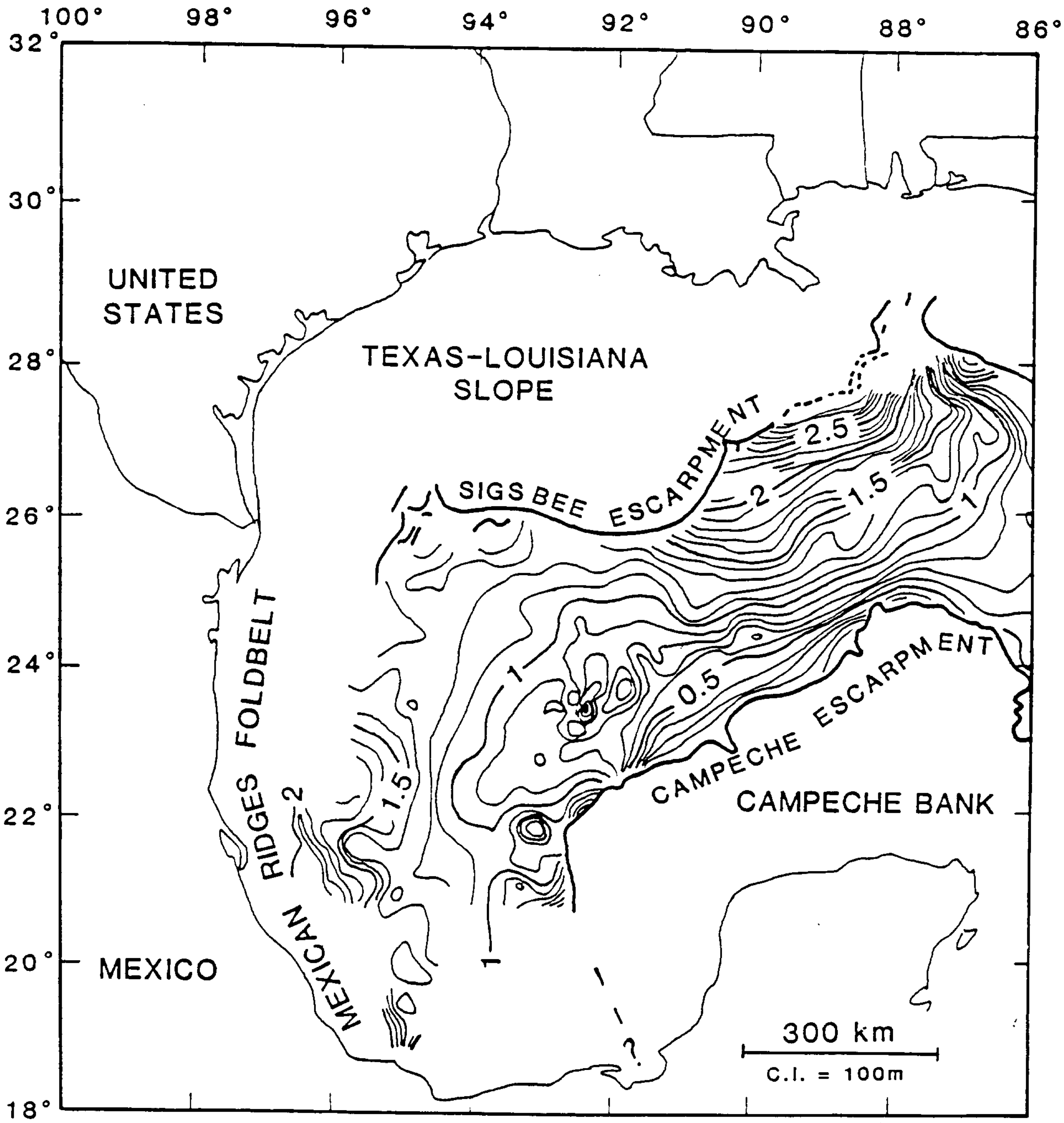


Figure 12. ISOPACH MAP OF MIDDLE TERTIARY (?) TO LATE MIOCENE UPPER MEXICAN RIDGES UNIT

After Shaub et al. (1984)

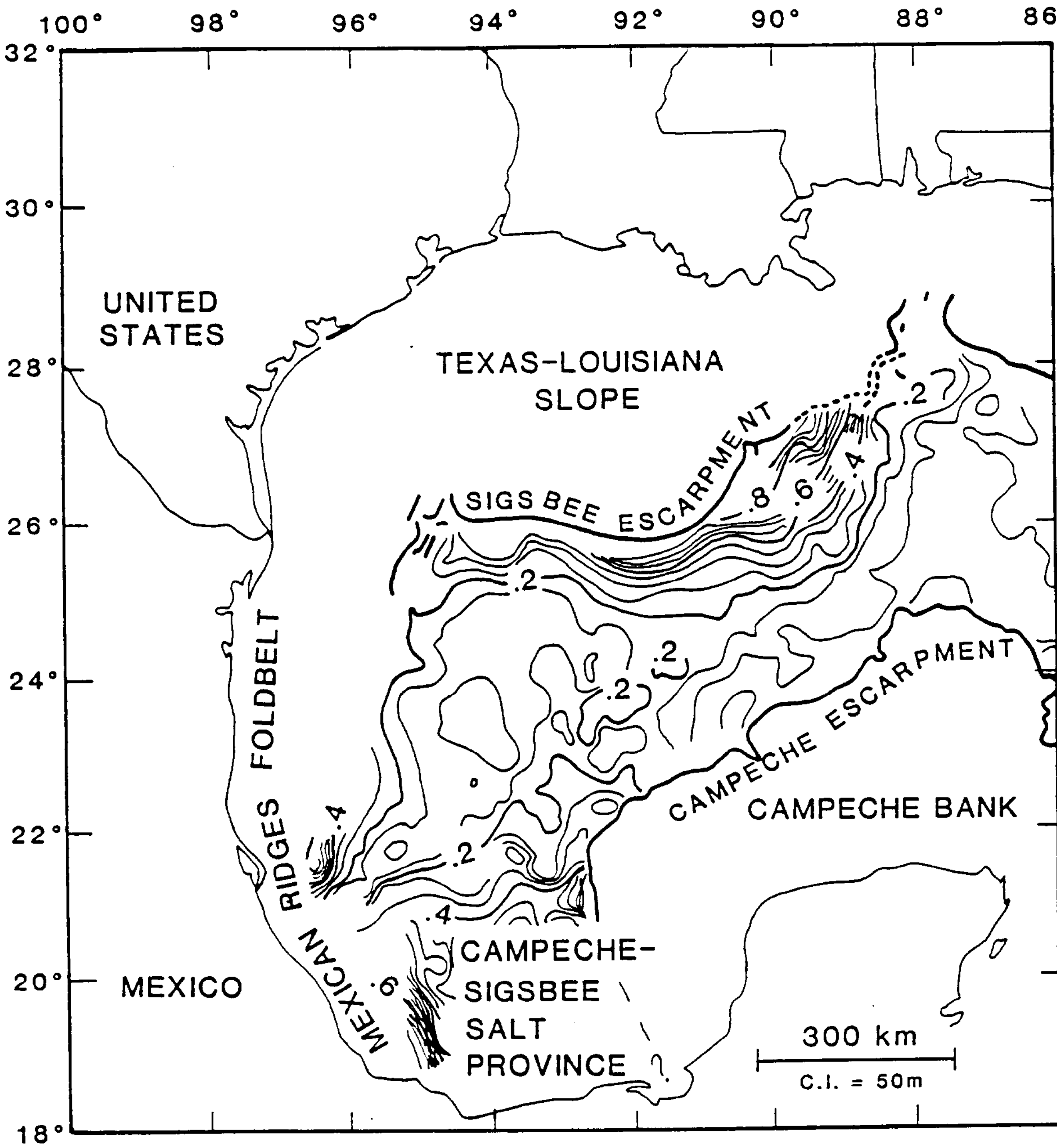


Figure 13. ISOPACH MAP OF LATE MIOCENE THROUGH PLIOCENE CINCO DE MAYO UNIT

After Shaub et al. (1984)

The Northwestern Margin of the Gulf of Mexico

The northwestern margin of the Gulf of Mexico includes the inland areas, coastal plains, continental shelf, continental slope of central and eastern Texas and western and central Louisiana (Figure 1). In terms of gas hydrate occurrences and potential, the continental slope is the only area of interest but its geologic evolution and present geomorphologic configuration can best be understood in a larger context. The salt tectonics was of paramount importance and is relevant to the presence of gas hydrate occurrences.

Structural Framework

Recent seismic refraction interpretations from the northwestern margin of the Gulf of Mexico (Buffler et al., 1980 and 1981; Ibrahim et al., 1981; Ibrahim and Uchupi, 1983) suggest the presence of so-called transitional crust between the well identified continental crust of the inner coastal plain and the oceanic type crust flooring the central deep part of the Gulf of Mexico (Sigsbee Abyssal Plain). The transitional crust, which appears to underlay the present Texas - Louisiana continental slope is very poorly defined because of perturbations in seismic wave velocities brought about by salt diapirism (Figure 6).

Seismic refraction velocity studies (Ewing et al., 1960; Antoine and Ewing, 1963) suggest that the thinned deep crustal basement of the syn-rift sequences is a sialic, crystalline, assemblage of late Paleozoic age.

Even within the modern concept of rifted continental crust, generally referred to as transitional crust, various authors have recently expressed different opinions about its constitution, width and boundaries for the northwestern continental margin of the Gulf of Mexico (Ibrahim et al., 1981; Ibrahim and Uchupi, 1983; Buffler et al., 1981; Hales et al., 1983; among others). The latter authors significantly depart from the present interpretation derived from Buffler et al. (1981), and support the idea of the thick Jurassic salt of the Texas - Louisiana continental slope resting on pre-salt sediments directly overlying oceanic crust.

Salt Diapirism and Basin Configuration

The geologic evolution of the northwestern margin of the Gulf of Mexico was controlled by the scale and extent of salt diapirism (Martin, 1976, 1978, 1980; Buffler et al., 1984).

The salt deposits throughout the northern Gulf of Mexico lie within roughly concentric, structural domains (belts), consisting of piercement (salt diapirs) and nonpiercement (salt swells or pillows) structures (Martin, 1978; McGookey, 1975; Woodbury et al., 1973; Figures 14 - 15). Martin (1978) has proposed the following terminology for the salt structure domains:

1. An inner domain consisting of the south Texas, east Texas, north Louisiana, and Mississippi salt basins, separated by paleohighs on which no or thin salt units were deposited: San Marcos, Sabine,

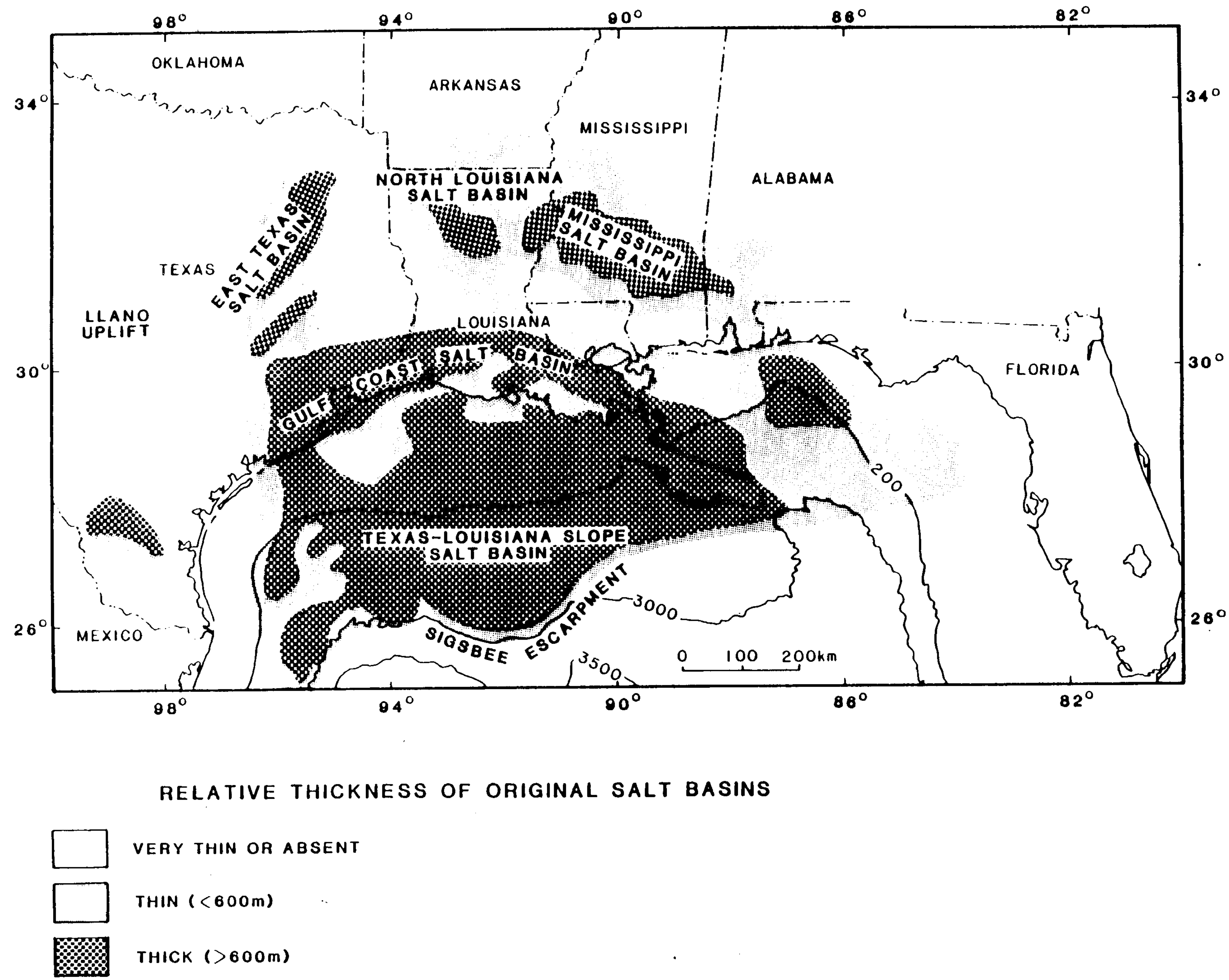


Figure 14. NORTHWESTERN MARGIN OF THE GULF OF MEXICO

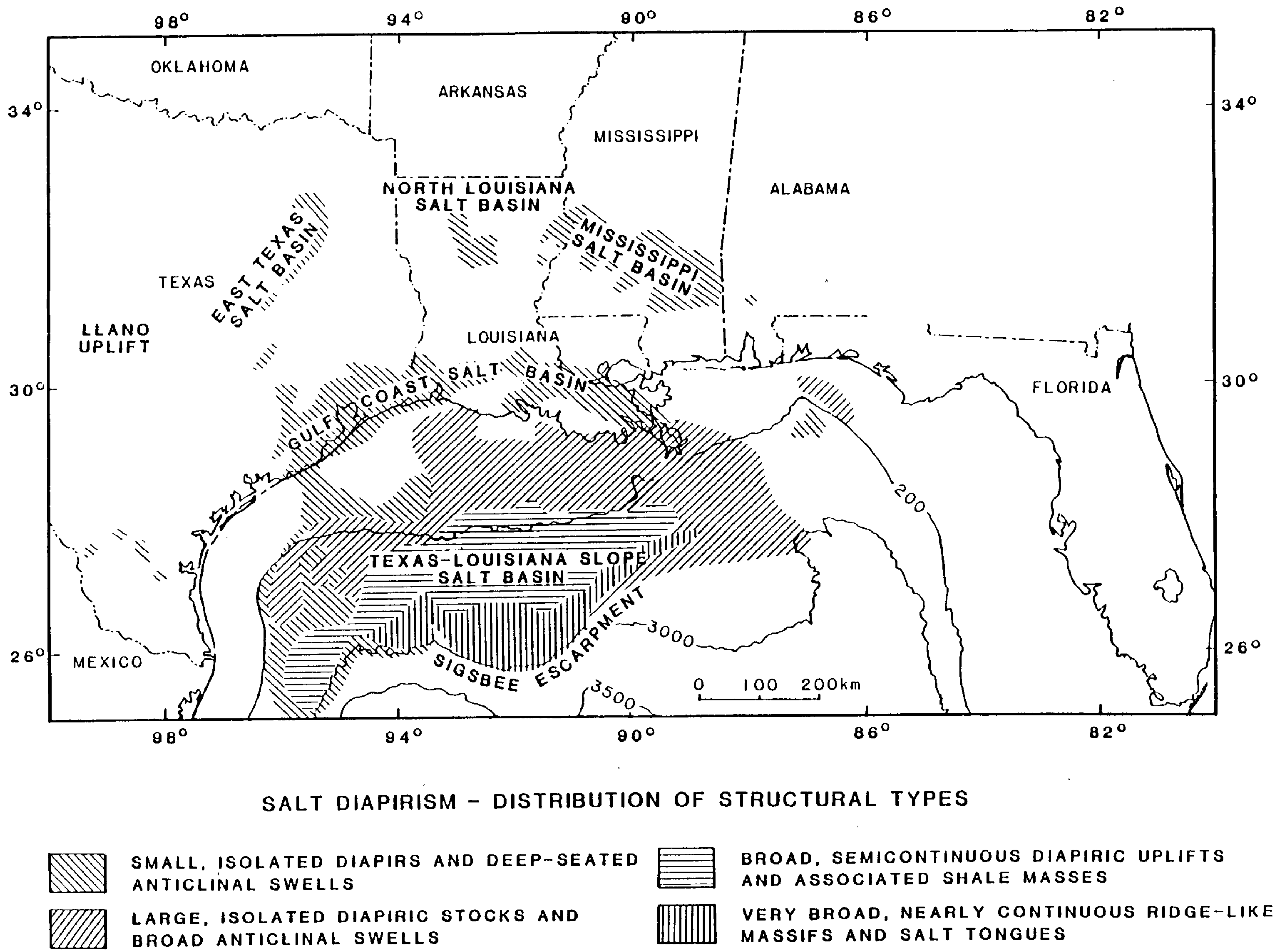


Figure 15. NORTHWESTERN MARGIN OF THE GULF OF MEXICO

After Martin, 1978

Monroe and Wiggins arches. The salt basins are characterized by small, high level diapirs and deep-seated anticlinal swells.

2. A middle domain, straddling the Texas - Louisiana coastal plain and inner continental shelf, called the Gulf Coast salt basin, also characterized by small, high level diapirs and deep anticlinal swells.

3. An outer domain that encompasses nearly all the Texas and Louisiana continental outer shelf and slope, typically called the Texas - Louisiana slope salt basin. This much larger diapiric domain can be subdivided on structural considerations as follows (Martin, 1973):

   A. The outer continental shelf, characterized by large isolated diapirs and broad anticlinal swells.

   B. The continental slope, characterized by broad semicontinuous diapiric uplifts and associated shale masses.

   C. The Sigsbee Escarpment at the outer portion of the continental slope, characterized by very broad, nearly continuous, coalescing ridgelike massifs and frontal thrust salt masses.

The broad outlines of these salt diapir domains presumably reflect the original configuration of the individual, syn-rift salt basins (Figure 14). The Gulf coast salt basins and the much larger Texas - Louisiana slope salt basin appear to represent two separate depositional entities or a complex subbasin system separated along most of its length by a paleohigh on which the salt is thin or absent by nondeposition. The increase of diapiric extent and activity and the concomitant amount of salt involved from the continental shelf to the Sigsbee Escarpment may or may not reflect a corresponding increase in original salt thickness (Humphris, 1978).

All recent investigators (Kupfer, 1974; McGookey, 1975; Martin, 1976 and 1978; Buffler et al., 1980 and 1981; Seni and Jackson, 1983; among others) agreed that the onshore and offshore salt basins correspond to a major evaporitic episode of late Early Jurassic age, presumably related to the rifting of the crust of the northwestern continental margin of the Gulf of Mexico. The evaporite sequences, constituted of massive salt (halite) and anhydrite, are estimated to have reached 500 to 1,500 m in thickness and are visualized to have been deposited in rifted grabens or half grabens which originated during the Late Triassic. The possibility of stratigraphic diachronous (time-transgressive) salt units within the separate basins has not been seriously addressed, in view of the inaccuracy of methods for dating salt strata. In the absence of data to the contrary, various salt sequences from the inland basins are regarded as one lithostratigraphic unit referred to as the Louann Salt. The offshore salt basin sequences are also considered stratigraphic equivalents of the Louann Salt. The Louann Salt from the inland basins is presently considered to be late Callovian to early Oxfordian (Imlay, 1980). This age determination is based on regional and long range stratigraphic correlations, in particular with the Minas Viejas Formation of northeastern Mexico and the overlaying Smackover Formation equivalent, the Zualaga Limestone. Historically the Louann Salt and its lithologic equivalents from the other salt basins was thought to be an equivalent of the Permian age Castile Formation of eastern Texas (Hazzard et al., 1945; Halbouty and Hardin 1956). Palynomorph identification by Jux (1961) and general stratigraphic relationships led to assigning a Late Triassic to Middle Jurassic age to the Louann Salt (Kirkland and Gerhard, 1971; Bishop, 1967; Imlay, 1980). The modern consensus, reviewed by Imlay (1980) dates the Louann Salt from the late Callovian to early Oxfordian (i.e. latest Middle Jurassic to earliest Late Jurassic age). The possibility of diachronous salt units within the separate subbasins should, however, be regarded as a likely possibility.

Pre-Quaternary Geologic Evolution

Pre-Late Jurassic. Pre-Jurassic strata of the inland areas of Texas and Louisiana do not outcrop to the surface. Their extent and thickness are inferred from subsurface data. The principal oil and gas reservoirs in the northwestern onshore margin of the Gulf of Mexico are the late Oxfordian Smackover Formation and the Cenomanian to Turonian Tuscaloosa Formation. Deeper units have consequently been very sparsely tested. Still, enough data for stratigraphic correlation and paleogeographic reconstruction is available from northeastern Texas, southwestern Arkansas, northern and central Louisiana, and southern Mississippi.

The basement of pre-Mesozoic age consists of rocks of the Ouachita fold belt and is composed of deformed Ordovician through Pennsylvanian, non to slightly metamorphosed lithologies (black phyllite, quartzite, and quartz-mucovite schist). The basement is unconformably overlain by a thick Late Pennsylvanian through Permian post-orogenic sequence composed of recurrent paralic clastic lithofacies, dominantly grey shales and fluvio-deltaic sandstones, and shallow water shelf carbonates. This sequence is lithologically similar to its counterparts in the mid-continent region north of the Ouachita Mountains (Meyerhoff, 1967; Vernon, 1971; Woods and Addington, 1973).

As much as 1,500 m of an irregularly distributed fluvio-deltaic red bed sequence is known to rest unconformably over Ouachitas assemblages and less commonly over the post-orogenic late Pennsylvanian to Permian sequence (Swain, 1949; Atwater, 1968; Kirkland and Gerhard, 1971; Vernon, 1971; Martin, 1976; Martin, 1977, and others). It is thought to have been deposited and subsequently preserved within a belt of peripheral graben and half graben whose trends parallel the arcuate shape of the former Ouachita fold belt. This fluvio-deltaic complex is dated, at least in part, as Late Triassic (Kirkland and Gerhard, 1971). Diabase dikes and basalt flows or sills are fairly common. On the basis of similarities in lithology and inferred tectonic and depositional settings, these red beds, named the Eagle Mills Formation, have been tentatively correlated with the Newark Group and equivalent formations of the Late Triassic to Early Jurassic rifted basins of the eastern U.S. and to the La Boca Formation of northeastern Mexico.

This red bed assemblage clearly represents the beginning of a syn-rift megasequence confined into extentional structures. It is unconformably overlain by a widespread thin sandstone and conglomerate unit which is capped by a massive anhydrite layer a few tenths of a meter thick in northeast Texas, southern Arkansas, and northern Louisiana. The anhydrite grades to a massive halite unit 500 to 1,500 m thick, known as the Louann Salt. It has been dated in one isolated instance (palynomorphs) as Late Triassic to Early Jurassic (Jux, 1961) but the present modern consensus, based on stratigraphic considerations is that the Louann Salt is Callovo-Oxfordian (Imlay, 1980).

The Louann Salt appears at a regional scale to have been deposited into fairly well individualized depocenters. The Louann Salt, and its anhydritic top member of southern Mississippi and Alabama is disconformably overlain by a thin clastic complex interpreted to have been deposited in an arid environment composed of Wadi type conglomerates, eolian sandstones and playa deposits, which grades upward to a regionally widespread, restricted marine to open marine, shallow water carbonate complex, the Oxfordian Smackover Formation. It is the deepest oil and gas reservoir from the northern Gulf coast, and it has been extensively investigated by the petroleum industry.

The Smackover Formation grades upward to, and interfingers with, a restricted, marine platform, clastic sequence composed of red and green shale, anhydrite, anhydritic shale, dolomite, limestone and salt. Various regionally persistent diachronous anhydrite units are reported from the base and the middle part of this sequence. In some areas this anhydrite is the dominant lithology. It can be reasonably construed as representing the waning stage of a widespread syn-rift sedimentary assemblage. This assemblage encompasses poorly documented regional facies changes and diachronous lithologic relationships. The recurrence of major evaporitic environments is particularly noteworthy. The consensus of published works on the Late Triassic to Jurassic of the inner Gulf coast is that an earlier continental depositional environment was dominated by arid climatic conditions and extentional tectonism which was replaced in early Late Jurassic time by a shallow to restricted, stable, marine shelf environment (Swain, 1949; Ewing and Antoine, 1966; Murray, 1966; Bishop, 1967; Halbouty, 1967 and 1979; Atwater, 1968; Kirkland and Gerhard, 1971; Newkirk, 1971; Braundstein, 1974; Jackson and Seni, 1983; and others).

Late Jurassic through Early Cretaceous. The Late Jurassic to Early Cretaceous strata of the northwestern Gulf coast represent the initiation of a new, post-rift stage in the evolution of the continental margin. They were deposited as thick differentiated, continental to marginal marine clastic wedges over a stable but steadily subsiding, shallow shelf setting which graded southward in south central Texas into a predominantly shallow marine carbonate environment (Rainwater, 1971).

Restricted tidal flats and lagoonal, hypersaline environments have been identified in the Early Cretaceous of the Mississippi embayment area; recurrent gypsiferous and gypsum-halite intervals are also documented.

The shallow marine carbonate environment of the Early Cretaceous of south central Texas was the precursor of the huge (Aptian to Cenomanian) carbonate platforms which almost completely fringed the Gulf of Mexico (Meyerhoff, 1967; Paine and Meyerhoff, 1970; Buffler et al., 1974; Figure 2). The Edwards Limestone of south central Texas and its subsurface extension known as Stuart City, represent the northern extension of the more developed carbonate platforms of eastern Mexico and the Yucatan Peninsula, but lacks the spectacular high constructional relief of these southern platforms. The Stuart City trend extends discontinuously northeastward through the Mississippi embayment where it is less conspicuously developed and interfingers with clastic shelf units joining the Florida carbonate platform to the east (Figure 16).

Late Cretaceous. The carbonate platforms of the northwestern rim of the Gulf of Mexico were progressively drowned by a Late Cretaceous marine transgression. The shallow water limestone was overlain by open shelf sandstone, shale, marl, and chalk. Local reefs and small carbonate platform buildups still occured locally. Paralic coal-bearing sandstones were deposited in some coastal areas (Holcomb, 1971). The Late Cretaceous was not characterized by clastic shelf progradation except for a brief period (Cenomanian to Turonian) in east Texas and Louisiana when a mud dominated slope system prograded beyond the Cretaceous reef trends. The deltaic shorelines reached the shelf margin and led to shelf edge progradation and gravitational instability. This transient shelf margin clastic complex, the Woodbine - Tuscaloosa Formations, was drowned and buried in post-Turonian time and a well defined clastic shelf edge was not reestablished until the Paleocene (Winker, 1982; Figures 16 - 17).

Tertiary. Early in the Tertiary, the northwest margin of the Gulf of Mexico began to receive large amounts of terrigenous material in response to the general uplift and erosion of large sections of the western interior (Laramide orogeny). The supply of sediments exceeded by for the rate of subsidence, leading to a generalized outbuilding and progradation of the shelf margin into the deep Gulf of Mexico. The general progradation pattern of the Tertiary and Quaternary was interrupted by short periods of regional subsidence and shifts in depositional patterns (Figures 16 - 17).

The encroachment of the clastic shelf over the deep basin reached as much as 400 km from the Cretaceous carbonate shelf. A figure of 250 km appears to be a representative average for the Tertiary and Quaternary progradation of the northwestern margin of the Gulf of Mexico. The major depocenters not only migrated in an offshore direction but also shifted through time in a clockwise fashion, from the Rio Grande embayment during the Paleocene to the Louisiana outer shelf in the Pleistocene (Woodbury, 1973; Winker, 1982).

Paleogeographic and facies studies have shown quite clearly that the type of depositional systems were rather constant through Tertiary time and bear strong similarities with the present Texas part of the Gulf coast depositional setting.

The detailed evolution of some of the major progradational cycles is quite complicated and difficult to unravel without adequate well coverage. The works of Fisher (1969), Fisher et al. (1970), Fisher and McGowen (1967), Edwards (1981), Berg (1981), Galloway et al. (1982), Jackson and Galloway (1984), and Winker (1982 and 1983) have documented the nature, extent, and permanence of the depositional processes which dominated the northwestern clastic margin of the Gulf of Mexico during the Tertiary time. The overviews of Lofton and Adams (1971), Tipswood et al. (1971), Shinn (1971), and Curtis and Picou (1980) present general approaches to the evolution of the U.S. Gulf coast.

The progradating shelf deltas were sometimes distinctly diachronous at a regional scale (Winker, 1982 and 1983). Periods of regional quiescence and marine reworking during transgressive episodes have been documented. The

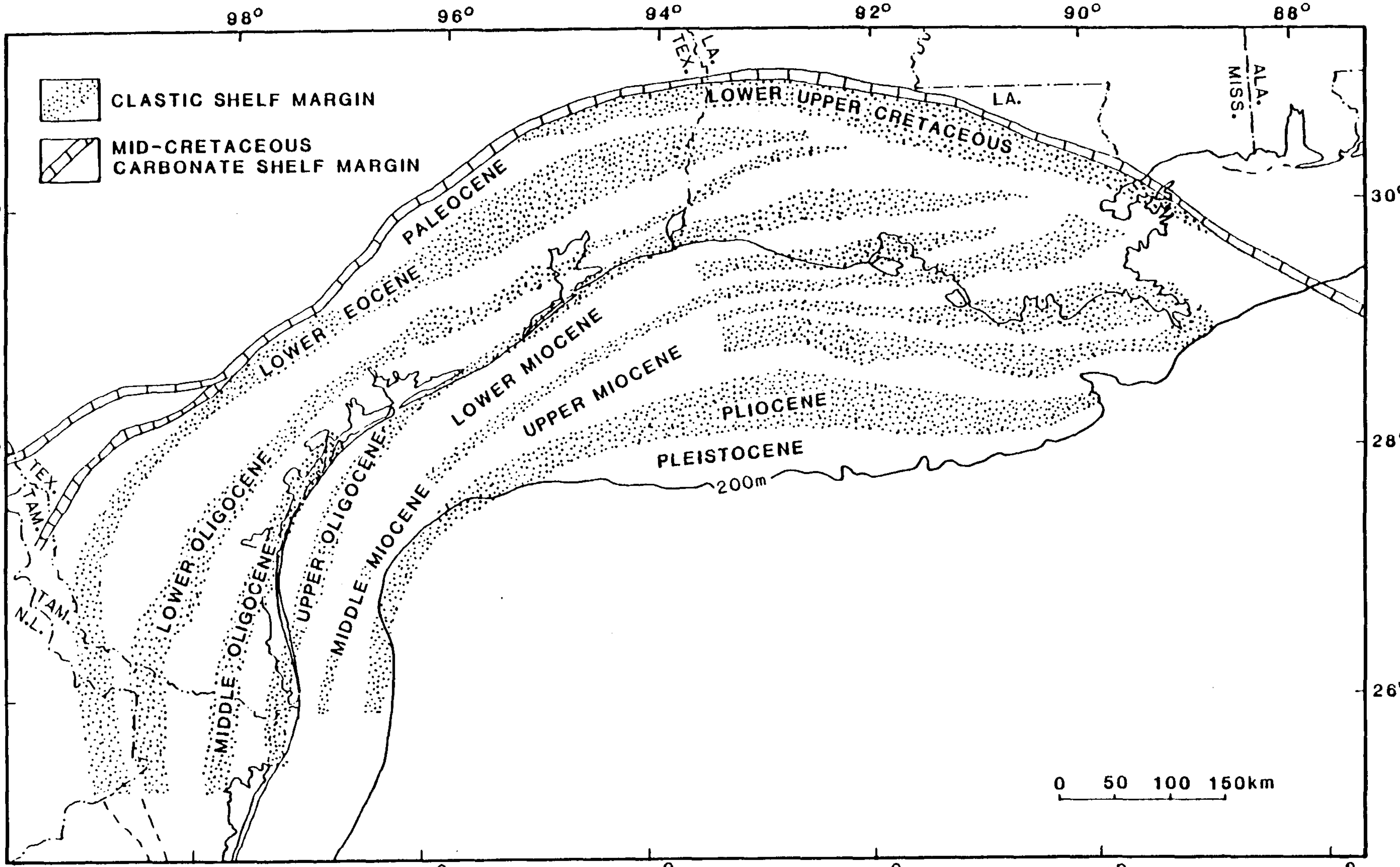


Figure 16. **LATE CRETACEOUS - TERTIARY PALEOMARGIN TRENDS OF THE GULF OF MEXICO**

After Winker, 1982

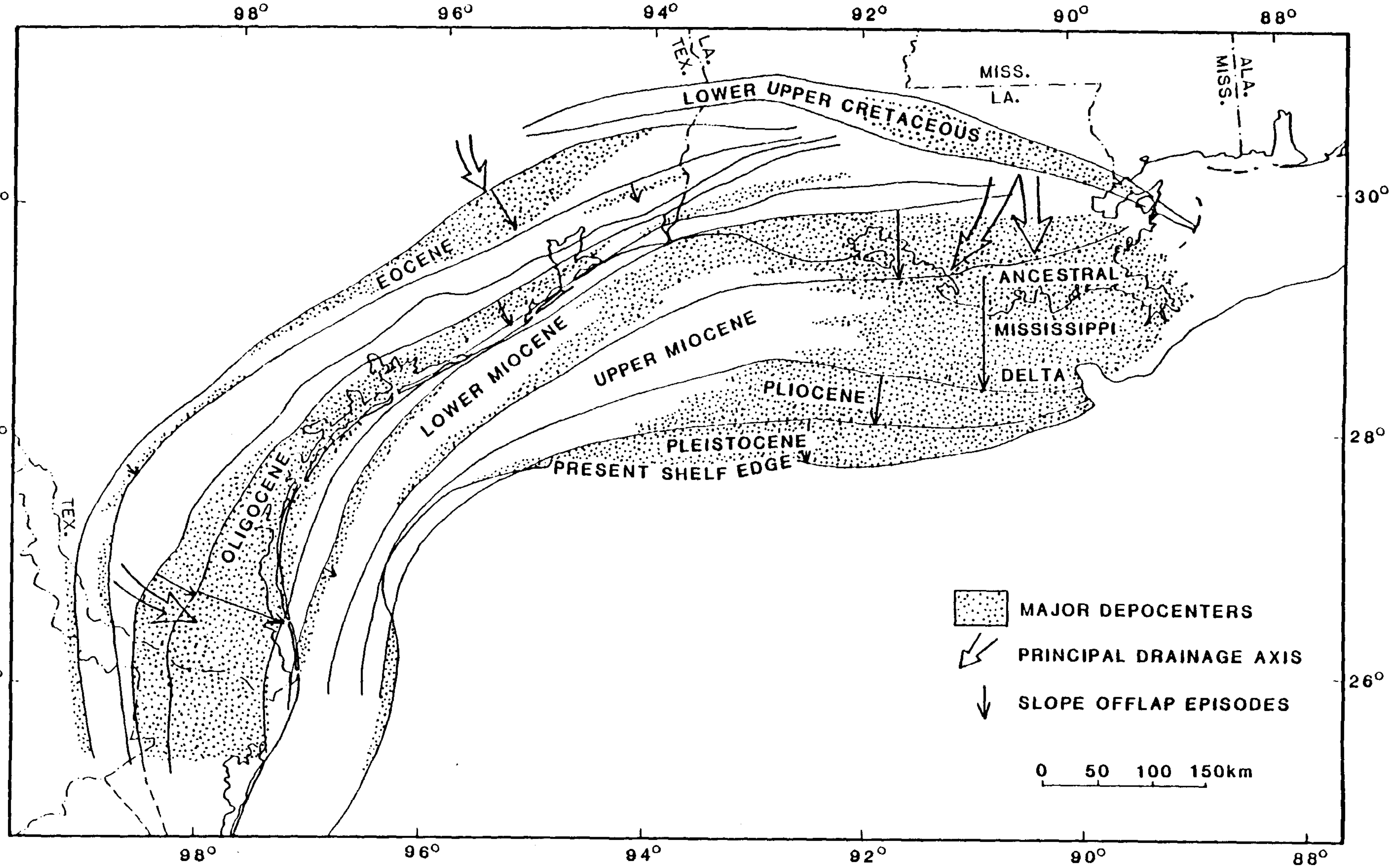


Figure 17. **POSITION OF THE MAJOR TERTIARY AND QUATERNARY DELTAIC CONTINENTAL MARGIN DEPOCENTERS**

After Winker, 1982

shelf margin retreated during the middle to late Eocene time and deposition did not readvance until the early Oligocene. The late Oligocene represents a period of marine transgression and limited shelf progradation. The early Miocene witnessed the shift of major depocenters from the southern Texas coast to the western Louisiana coast, apparently representing the establishment of the Mississippi drainage as the largest sediment source for the Gulf Basin (Winker, 1982). From early to late Miocene, the ancestral Mississippi depocenter migrated eastward to southeastern Louisiana. This pattern was reversed in the late Miocene to Pliocene; the Mississippi shelf margin depocenter migrated westward toward central, west central Louisiana (Powell and Woodbury, 1971; Woodbury, 1973; Figure 17). Fluvial and deltaic systems, normal to the paleoshoreline and coastal systems fringing the paleoshoreline prograded over shelf sands and delta front shelf margin mudstones.

On a regional scale the rapidly subsiding shelf margin acted as a major sediment trap permitting the accumulation of hundreds to thousands of meters of shallow water deposits during a major progradational (regressive) episode. Rapid outbuilding of the shelf margin over the continental slope led to a steepening of the shelf edge and sediment instability. Typically the shelf margin was dominated by subaqueous gravity depositional processes. Sliding, slumping and growth faulting were predominant mass wasting sedimentary processes overshadowing other downslope sediment transport processes such as turbidity current, mudflow and shallow slump (Jackson and Galloway, 1981; Winker, 1982 and 1983; Edwards, 1983).

The rapid influx of shallow-water shelf margin deltaic sands and their dumping over unconsolidated slope mudstone led repeatedly to overloading of the mudstone units and their gravitational failure. Shale diapirism and syn-sedimentary listric faulting (growth-faulting) and associated antithetic faulting are conspicuous structural features directly related to slope failure. This rapid shelf margin deltaic progradation over unconsolidated slope mudstones and the accommodation by growth faulting of this large material influx resulted in accumulation of sediment at a rate at least ten times that for the corresponding deltaic shelf sequence. The recurrent pattern of this basic depositional process led to repeated growth faulting and shale diapirism and to the curious shale ridge configuration observed in many distal portions of shelf edge prograding clastic systems. Growth faults and shale diapirism in many cases were not able to alleviate the consequences of rapid burial and progradation. Insufficient time for dewatering and degassing of the slope mudstone led to geopressurization, compounded by hydraulical isolation of shallow water sandstone bodies juxtaposed against slope mudstones by growth faulting. While gravity sliding of the continental slope created an extensional regime along the shelf margin, the corresponding compressional regime along the lower slope was important in initiating salt and shale diapirism. Rapid sediment accumulation over the slope led to increase in confining pressure and upward flowage of mudstone, and more importantly the mobilization of deep salt units. The progradation of the shelf margin over these salt structures greatly complicated the style of growth faulting and depositional processes.

Quaternary Geologic Evolution

The geologic evolution during Quaternary time, particularly the last period of lowstand of sea level, had a direct impact on the sedimentary processes and submarine topography. Even more so than during the late Tertiary, the Quaternary continental margin was strongly differentiated into shelf and slope regions which will be treated independently. The development of the continental slope, from which at least eight samples of gas hydrates have been recovered (Brooks and Bryant, 1985a) and the continental shelf are interrelated, especially when the influence of the eustatic fluctuations of sea level is taken into consideration.

Continental Shelf. The Texas - Louisiana continental shelf extends from the Mississippi River Delta westward to the Rio Grande where it adjoins the east Mexico shelf along an approximate boundary at 26°N latitude (Figure 1). The width of the shelf varies from about 100 km off the Rio Grande to more than 200 km south of the Texas - Louisiana state boundary near 93°30'W longitude. The eastern portion of the shelf is covered by the Mississippi Delta whose presently active lobe almost entirely crosses the shelf southeast of New Orleans and empties its sedimentary load directly onto the continental slope. The break in slope that marks the outer edge of the shelf throughout the region is difficult to define but, on the average, is about 120 m. The surface of the shelf is generally smooth, but locally marked by subdued topographic features including relics from times of glacially lowered sea level. Low fault scarps and small banks and mounds, many of which express the presence of underlying salt or shale diapiric structures are also found on the shelf. Calcareous banks are the most prominent features of this rather smooth shelf.

The present modern shelf is unsuitable for the formation and preservation of gas hydrates due to its depth and bottom water temperatures, but its Pleistocene evolution has a direct bearing on the depositional and structural evolution of the continental slope.

Main Geologic Features. Pleistocene construction of the northwestern continental margin of the Gulf of Mexico is the result of upbuilding and outbuilding of the continental shelf. The Pleistocene witnessed a continuation of the basic sedimentary processes which dominated the Tertiary with important complications brought about by the fluctuations of sea level. The general seaward progradation of a clastic shelf edge apron continued unabated in response to the sediment influx of the ancestral Mississippi River system. The area of maximum deposition, following the middle Miocene westward depocenter shift from southeastern Louisiana to west central offshore Louisiana was preserved during the Quaternary (Figure 17; Woodbury et al, 1973; McGookey, 1975; Valentine and Poag, 1976; Winker, 1982) until very recent time when the Mississippi Delta shifted its main course to eastern Louisiana.

These developments occured in conjunction with the fact that significant amounts, possibly a majority, of the the sediments supplied by the ancestral late Pliocene to Pleistocene Mississippi River, were directly diverted to the deep-water continental slope and rise and built the Mississippi Deep Sea Fan (Figures 12 - 13).

Major progradational or outbuilding phases of deposition occured during periods of lowstands of sea level with a net overall result of major enlargement of the continental shelf and the accumulation of over 3,500 m of sediment along the outermost shelf and upper continental slope. As during the Tertiary, rapid shelf progradation brought instability; widespread gravity-induced slope failures led to resedimentation of large amounts of sediments down to the continental slope and formation of a lithologic environment conducive to gas hydrate formation.

Late Pleistocene Margin Deltas. Glacial eustatic fluctuations of sea level have been one of the major factors in developing the morphology and sedimentation processes of the continental margin of the northwestern Gulf of Mexico. Repeated drops of sea level coupled with high sedimentary supply have resulted in the deposition of a thick sedimentary wedge and ensuing progradation at the shelf edge over the upper continental slope (Frazier, 1974; Sidner et al., 1977 and 1978; Suter and Berryhill, 1985). Spillover beds with dips as great as 5° are documented from the shelf edge region from a broad continuous belt up to 30 km wide of prograded, regressive sediments (Lehner, 1969; Figure 18).

Ancient fluvial systems were eroded into the subaerially exposed shelf during periods of low sea level, channeling very large amounts of sand and mud to the shelf edge, thereby leading to the formation of the shelf margin, fluvially dominated deltas, which outbuilt over the steeper continental slope.

Only the latest Wisconsin shelf margin lowstand delta complex has been documented in detail; five distinct delta systems have been identified, sometimes with their attendant shelf fluvial channels (Suter and Berryhill, 1985). Stream channels, lagoons and barrier beaches are common late Wisconsin relics across the modern shelf (Winker, 1980 and 1982; Martin and Bouma, 1978; Suter and Berryhill, 1985). Similar features related to former periods of sea level lowstands have been obliterated or prograded over and have not been recognized. In fact, only one pre-late Wisconsin shelf margin delta system has been firmly identified (Sidner et al., 1976 and 1978). Seismic stratigraphy analyses document two main superposed prograding sequences (Stuart and Caughey, 1977). The first episode corresponds to a period of prolonged sea level lowstand at 350,000 y.b.p. and the subsequent continued progradation during a period of slow paced, sea level rise. The second phase of shelf edge outbuilding is known from the middle late Wisconsin and is inferred to have begun around 80,000 y.b.p. Its latest phase (late Wisconsin, 18,000 to 10,000 y.b.p.) is the best preserved and the best understood. The transgressive, open shelf deltaic deposits are not as well developed as in the first Wisconsin cycle, a characteristic presumably related to the rapid rise of sea level which succeeded the last glacial event. Nonetheless, some fluvial, nearshore and barrier bar relict deposits are known.

The late Wisconsin shelf margin prograding phase is, due to its preservation, the most amenable to precise and comprehensive study (Suter and Berryhill, 1985). Five main shelf margin deltas have been recognized from the Rio Grande to the central Louisiana shelves with their attendant large buried fluvial shelf networks and their prodelta slope environment. The mapping of the paleofluvial networks is the key for the localization of the shelf margin deltas, and it indicates that drainage patterns of the exhumed shelf during the last lowstands of sea level were considerably different from those prevailing now. In particular the course of the western distributaries of the ancestral Mississippi River was drastically offset to the west. River channel diversion was an active process throughout the late Wisconsin regression with upland channel cutoffs being augmented by shifts in river course caused by diapiric movement.

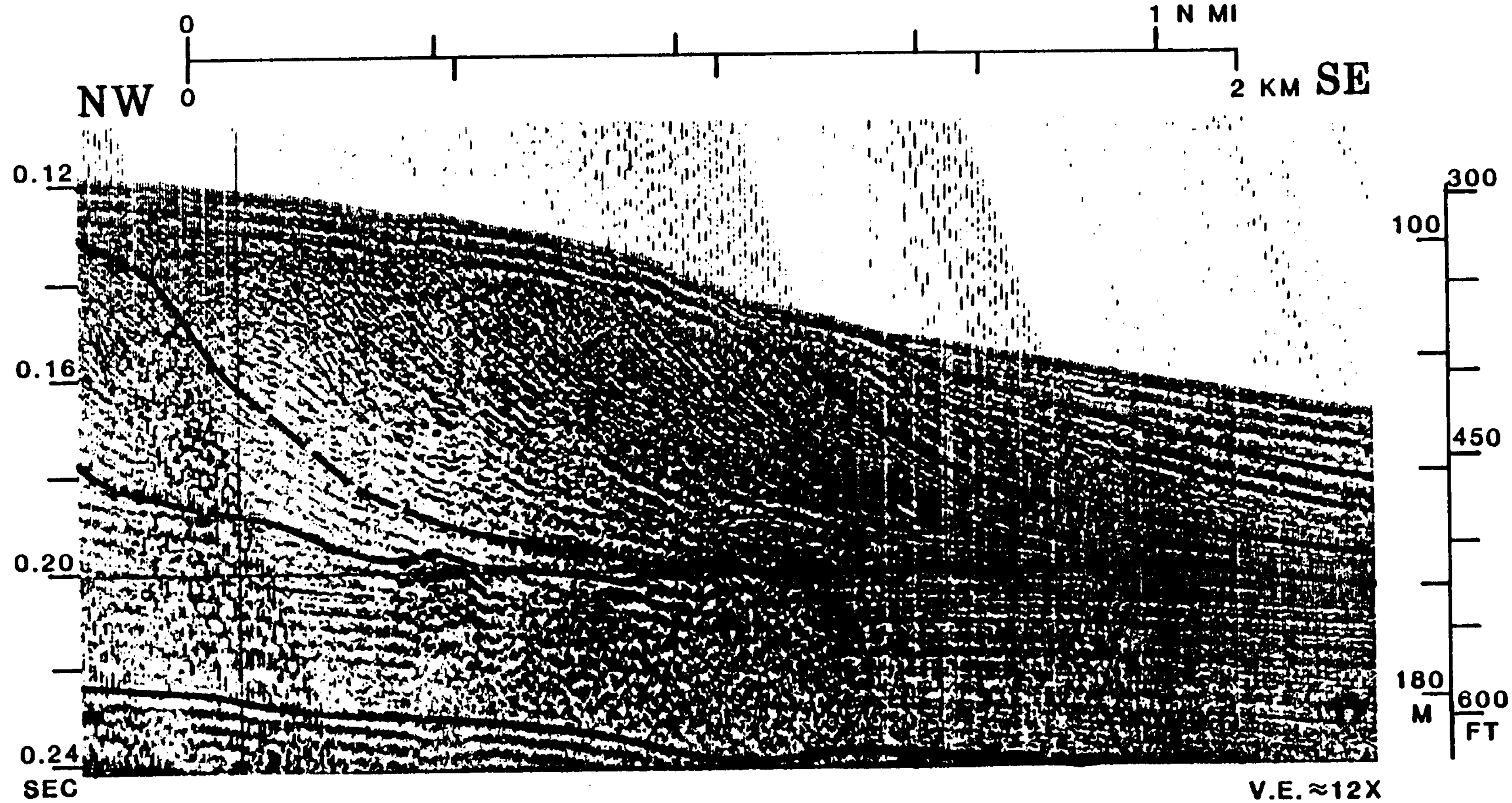


Figure 18. SEISMIC PROFILE THROUGH FORESET BEDDING OF A LATE PLEISTOCENE SHELF MARGIN DELTA

After Bouma, Stetling and Feeley in Bally (1983)

These late Wisconsin shelf margin deltas range in areal extent to more than 5,000 km^2 and to more than 180 m in thickness. These fluvially dominated deltas have an internal geometry and shape constrained by their position at the shelf margin. Waves and longshore currents redistributed the clastic material parallel to the shelf edge, thereby leading to an elongated arcuate to lunate pattern. These shelf margin deltas are multilobate both vertically, suggesting sea level fluctuations, and laterally, suggesting river mouth switching.

Due to the direct progradation over the much steeper gradient of the continental slope, these deltas could not prograde in a pattern akin to the present Mississippi Delta. The newly created delta slope was steep and unstable and consequently prone to widespread and repeated sliding, slumping, and various subaqueous gravity processes, bringing large amounts of clastic material into the lower slope environment.

In the case of the Mississippi River late Wisconsin shelf margin delta, thick sequences of slumped sediments are largely, but not entirely confined to two ancient submarine troughs that predate the last lowstand of sea level and may have existed during the last three lowstands. The troughs were avenues of transport for large volumes of sediment to the lower continental slope and may have been formed by retrogressive head canyon slumping. Much of the fill is slumped sediment overlain by a relatively undeformed prograded sequence. The toe of these slump complexes abutted in some instances against the flank of diapiric structures. Fault offsets, offlapping and truncational relationships testify to the post-slumping diapir rise. In one documented instance, a fan downslope of a shelf margin delta was damned by a rising diapiric barrier.

Seaward progradation was rapid; extensively slumped upper slope sediments were overridden by younger shallow-water deltaic deposits. A pattern of overlapping slide and slump masses and deltaic sands, followed by fluvial and nearshore barrier island sediments deposited under the reworking influence of slowly rising sea level continues for more than 20 km in some instances.

The shelf margin deltas were very important for dispersal of sediment in the upper continental slope. The vast majority of the sediments deposited were clays and silty clays, with minor amounts of sands. Shelf margin deltaic sands, submarine fan facies, and major slide deposits could accumulate simultaneously in close proximity in this type of tectonic setting, even within the same interdiapiric basin of the upper continental slope, a sedimentary environment favorable for gas hydrate formation.

Late Pleistocene Isostacy. The dynamics of the shelf margin, upper continental slope region was complicated by crustal flexure. The indirect evidence for postglacial isostatic downwarping (and/or accelerated tectonic subsidence) has been reviewed by Poag (1973). At least twenty-six distinct submerged surfaces ranging from 2 to 223 m below present sea level have been recognized for the late Wisconsin. A drop of sea level as low as 223 m below present sea level is approximatively 100 m lower than the generally accepted lowstand during maximum Wisconsin glaciation. Moreover, Lehner (1969) has reported a number of truncated diapiric sea knolls whose crests are as deep as 533 m below sea level and suggested that it was the result of postglacial downwarping of the upper continental slope. The work of Fisk and

McFarlan (1955), and others, documented even more thoroughly the influence of considerable downwarping associated with deposits beneath the ancestral Mississippi River.

Such rapid and extensive change in relative sea level may affect gas hydrate stability. Such water depth changes in the Gulf of Mexico would cause a vertical shift in the zone of gas hydrate stability due to a change in pressure, perhaps permitting concentration of hydrates to amounts greater than those possible in conditions of static water depth. The effect of large-scale rise and fall of sea level on gas hydrates would be comparable to that of large-scale submarine slumping and resedimentation which was explored by Krason and Ridley (1985a, 1985b). According to that interpretation, a repeated vertical migration of the zone of gas hydrate stability allows free gas to be liberated from hydrates dissociated when the zone of hydrate stability rises. This gas is incorporated in a more concentrated form into hydrates when the stability zone moves down the sediment column.

Continental Slope. The Texas - Louisiana continental slope is a 120,000 km^2 region of complex and hummocky submarine topography. The slope, with an average gradient of 1°, spreads westward from the Mississippi fan to the pronounced bend in the continental margin where it merges with the Rio Grande continental slope. The continental slope is bounded seaward by a pronounced steepening (up to a 10° slope) known as the Sigsbee Escarpment or Sigsbee Scarp (Figure 1).

Salt Diapirism. The hummocky topography which is characteristic of the Texas - Louisiana continental slope is caused by salt diapirism at an unprecedented scale (Figure 2). The Sigsbee Escarpment represents the physiographic expression of the lobate frontal edge of this diapiric province and is cored in its entire length, estimated at 450 km, by a massive composite system of salt ridges (Martin, 1973; Garrison and Martin, 1973; Martin, 1976 and 1978; Amery, 1978; and others; Figure 19). The entire Texas - Louisiana continental slope province is underlain by broad, semicontinuous, diapiric salt structures which interconnect at relatively shallow subbottom depth.

Salt structures can be grouped according to structural relief, size, and overall density, defining morphological belts of different structural maturity. Salt structures on the uppermost continental slope (between 200 and 600 m water depth) are commonly small, pluglike masses having diameters of 7 to 13 km and generally rising 2,000 to 3,000 m above a common base level of interconnection. Relief on the salt surface in the middle slope region in 600 to 1,400 m water depth is about the same as that for the upper slope diapirs. However, structures there are generally elongate ridgelike masses, with northwest to southeast through north to south trends. In profile views these structures appear as flat-topped, steep-flanked coalescing massifs separated thick sedimentary interdiapiric basins. The salt surface under the lower continental slope is constituted of large coalescing swells, 28 to 46 km across which rise about 1,000 to 2,200 m above their bases, trending northeast to southwest and separated from one another by broad structural depressions (synclinal basins of Lehner, 1969, and others), filled nearly brimfull with sediments. The salt structures of the Texas - Louisiana continental slope terminate abruptly a the giant, steep, salt front underlying the Sigsbee Escarpment. This geologically rather unique salt wall is made up of accreted elongated salt ridges. Allochthonous salt masses ("salt tongues") thrust over Pleistocene sediments for distances up to 12 km have been observed from seismic profiles (Amery, 1969; Buffler et al., 1969).

The overall pattern and size of salt diapirism under the Texas - Louisiana continental slope and its frontal portion (Sigsbee Escarpment) reflects the rapid sediment loading of the area. Continuous encroachment of the continental slope clastic wedge into deep water triggered frontal salt mobilization and uprising. The salt masses of the Sigsbee Escarpment may represent vertical salt ascent structures compounded by additional salt squeezed from beneath the advancing clastic wedge. Alternatively, they may be large allochthonous lobes of salt thrust into or over late Cenozoic sediments from distances of 50 to 75 km by regional lateral flowage under an advancing sedimentary apron (Humphris, 1978 and 1979; Buffler et al., 1981; Figure 19). The bulge configuration of the lower slope, the massive lobate structure of the Sigsbee Escarpment, and the occurrence of occasional, faint, flat lying seismic reflectors beneath the salt all support the lateral flow hypothesis.

Most of the recent detailed published studies covering specific areas of the Texas - Louisiana continental slope and recent advances were made by Bouma as major investigator. Most of what follows is derived from Bouma et al. (1978), Bouma, Martin and Bryant (1980), Bouma et al. (1981), Bouma (1983), and Bouma, Sterling and Feeley (1983). Additional material is taken from Garrison et al. (1977), Woodbury (1977), Woodbury et al. (1977 and 1978), Sangree et al. (1978) and Tatum (1979).

Only a few salt diapirs of the upper continental slope have actually been cored (Lehner, 1969). It appears that most of the high level, steeply flanked and flat topped salt bodies are mantled by mudstone and some of the diapiric structures may consist entirely of mudstone ('shale diapir'). Locally, the mantling mudstone and, in a few places, the salt or caprock are probably exposed at the sea floor. However, most of the diapiric bodies are covered by late Cenozoic sediments that range in thickness from a few meters to several hundred meters. The diapiric growth is so dense that diapirs can delineate enclosed depression at the sea floor know as intraslope basin (type II of Bouma). The slope and evolution of these intraslope basins are directly related to the diapiric growth of the flanking salt structures. These intraslope basins are very important in terms of hydrate occurrences.

Late Pleistocene Isostacy. Truncation of sea knolls whose present depths increase in an offshore direction down to a maximum of 533 m below present water depth, in conjunction with dead corals and dolomitic boulders dredged from a diapir knoll are supporting evidence for a local or regional postglacial downwarping and/or accelerated tectonic subsidence of the present upper continental slope in late Wisconsin time (Lehner, 1969; Poag, 1973). This phenomenon introduces an extra parameter in the crustal dynamics of the upper continental slope.

The irregular topography of the Texas - Louisiana continental slope is maintained by the continuous upward motion of the salt diapirs. The diapiric motion is perpetuated at a regional scale by the general shelf clastic progradation and at a local scale by the increased sediment load exerted by the intraslope basin infill. This sediment infill is itself largely due to massive slumping and thinning of the sediment cover from the unstable flanks of the diapirs. This dynamic system appears to be in a state of disequilibrium, and

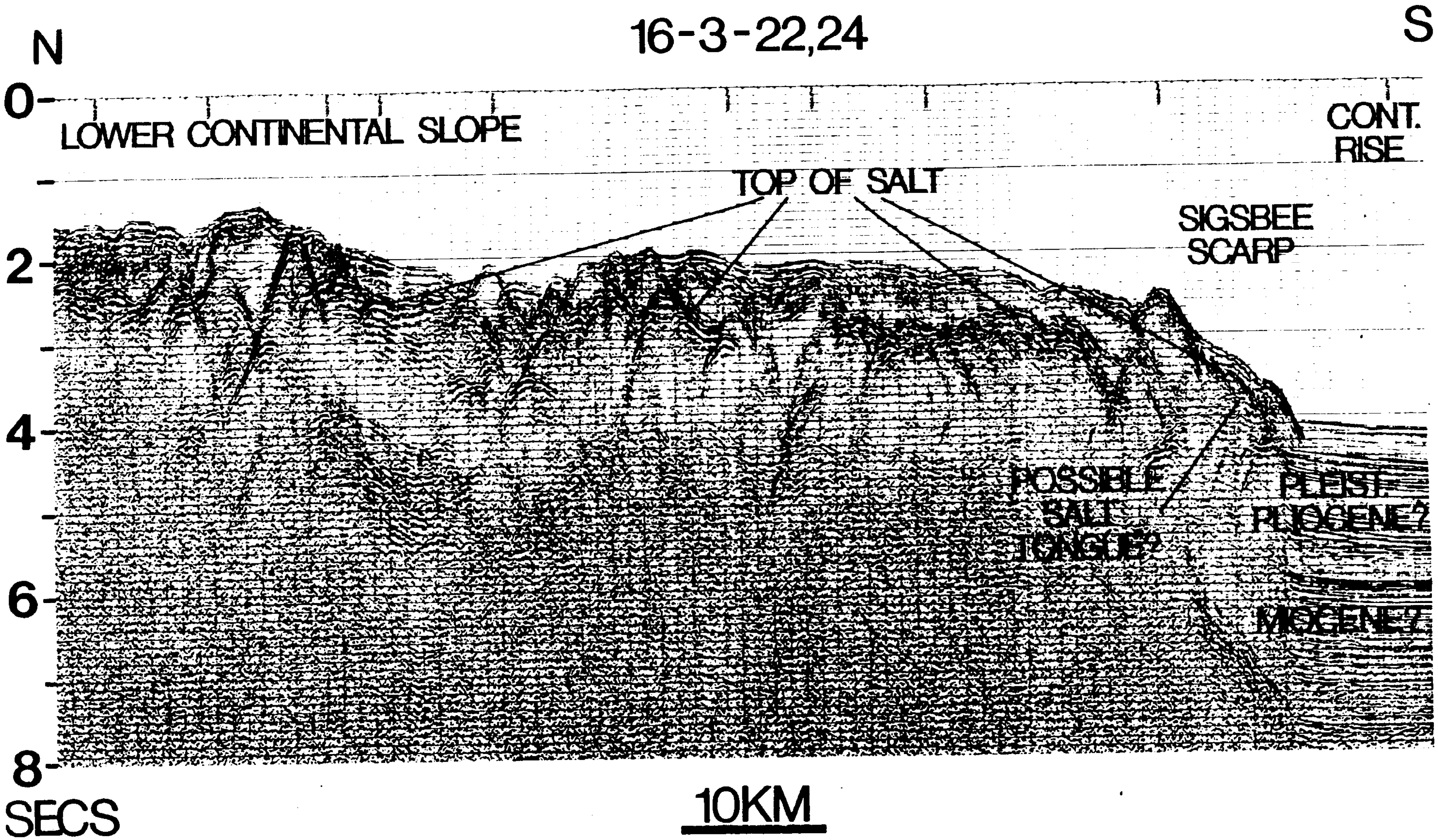


Figure 19. SEISMIC PROFILE THROUGH THE SIGSBEE ESCARPMENT

After Buffler, In Bally (1983)

it leads to widespread slope unstability and very complex, transient, bottom and subbottom geologic configurations.

Depositional Systems. The Texas - Louisiana slope is not transected by any permanent, major submarine canyon system (Bouma, 1983b). Only a limited number of submarine canyons appear to exist and their continuation through time is not certain. Their overall course is directly related to a fairly constant supply of sandy material from the shelf edge. These canyons are conduits for shelf sand to deep-water environments. They ultimately discharge their sedimentary load in the lower slope and continental rise region, terminating in small deep sea fans which coalesce to form a narrow clastic belt at the foot of the Sigsbee Escarpment, over the continental rise. A large amount of material is blocked on its downslope journey by rising diapir barriers which are breached if the canyons do not find a diversion route. Only after the ponded canyon fill reaches spill point level does the canyon become active again as a sediment conduit. The diapiric ridge blocking can be so intense as to break apart various portions of a canyon and to segment the former depositional system into disconnected, enclosed depressions (intraslope basin, type I of Bouma). At the distal portion of the canyon, the Sigsbee Escarpment continued frontal seaward encroachment over the continental rise by successive seaward formation of new salt ridges. This led to the fragmentation of the deep sea fans and their incorporation into individual, disconnected sea floor depressions within the diapiric province of the Gulf of Mexico northern continental slope.

The majority of the sediments of the continental slope are fine grained mud and silty mud deposited by various gravity induced mass-washing depositional processes and by hemipelagic settling. The sediments are not related with distinct dispersal systems.

The extent of late Wisconsin to present large-scale slumping along the upper continental slope has been dramatically documented by Lehner (1969). Slides, detached from the shelf edge for a downslope displacement of 55 km have been shown to have come to rest in mid-slope in 900 to 1,200 m of water. Older slump fills (Wisconsin) have apparently been traced across the lower slope for 165 km along strike. The extent of large-scale sliding, slumping and general gravity induced mass movement is considerable; Woodbury et al. (1978) have estimated the extent of recently displaced sediments from the shelf edge as reaching as far as 90 km downslope.

Pleistocene Sea Level Variations. During periods of relative lowering of sea level, very large quantities of sand and mud are transported from the continent and the inner shelf across the broad, newly exposed shelf to the water's edge. There it is deposited in rapidly prograding shelf margin deltas and/or redistributed along the shelf edge and upper continental slope by longshore currents (Suter and Berryhill, 1985). Following the initial outbuilding stages, large amounts of material are stored temporarily in constructional shelf edge - upper slope prograding complexes (Figure 18). Gravity instability becomes predominant and sets in motion mass wasting mechanisms. Slumps and turbidity currents bring vast amounts of clastic material into the lower slope environment. The rate of sea level lowering appeared to have been rapid during the late Wisconsin, thus accelerating these processes. Once the relative lowering of the sea level slows down and ultimately stops, the shelf rivers can no longer bring large amounts of material to the coastline, and sand is replaced by mud, which furthers the shelf edge instability (Bouma, 1983). The succeeding rise in sea level and the water high stand create a wide submarine shelf and makes it impossible to move sand to the upper continental slope. Bottom currents and hemipelagic sedimentation deposit a blanket of stratified mud over the sea floor of the continental slope. Diapiric upward motion leads to slope failure and the slumping of not only hemipelagic muds deposited over diapiric hillocks but also of the sand bodies deposited during low stand intervals.

High resolution, shallow penetration seismic records clearly show a cyclic pattern in slope sediments that can be explained by the interplay of the mechanisms just mentioned, i.e. thick, slump and turbidity current deposited clastics during period of low stands and thin hemipelagic draping muds during period of high stands.

Intraslope Basins are geologically short-lived, localized depocenters whose formation and subsequent evolution is directly controlled by salt diapirism. A large number of gas hydrate occurrences have been reported from these intraslope basins.

The upper and middle continental slope of Texas and Louisiana is characterized by randomly distributed, steeply flanked diapiric stocks (3 to 30 km across) and ridges ('salt anticlines') up to 50 m in length. These thinly covered salt bodies are separated by deep enclosed topographic depressions containing thick sections of stratified deposits. Three basic types of intraslope basins have been delineated (Bouma et al., 1978; and Bouma et al., 1980).

The first type, labelled blocked canyon intraslope basins, developed as previously explained by salt damming of a submarine canyon, thereby forming an elongate enclosed basin. Sand bodies, resulting from the pre-damming history of this section of a canyon form the lower sedimentary infill of this type of subbasin, and are overlain by muddy sediments deposited after the sand supply was cut off (Figure 20). A number of gas hydrate occurrences have been recorded recently from this type of depression from the Green Canyon Lease area (Brooks and Bryant, 1985a).

Diapiric uplift in noncanyon areas often results in the formation of interdomal intraslope basins by coalescence of individual salt masses. Such basins, of very irregular forms do not contain appreciable amounts of sand. Bottom water in these basins may be oxygen deficient. A spectacular example, filled with a 200 m deep hypersaline, anoxic brine has been documented from the Louisiana lower continental slope. This depression, named the Orca Basin, contains disseminated gas hydrates in its bottom sediments (Brooks et al., 1984; Figures 21 - 24).

A third type of intraslope basin, collapse basins, has also been reported, although this type of basin seems to be rather uncommon. They are formed by normal faulting over the crest of salt diapirs and represent tensional grabens or collapse structure associated with subfloor salt dissolution (Figure 25). They are less well studied than the first two intraslope basin types and no gas hydrate occurrences have been reported from them. Thus, the following comments regarding the evolution of intraslope basins will directly concern the first two types.

The sediment fill of an intraslope basin typically displays an alternation of slumped masses, grain flow and turbidity current deposits interbedded with hemipelagic sediments. Both types of sequences appear to be related as

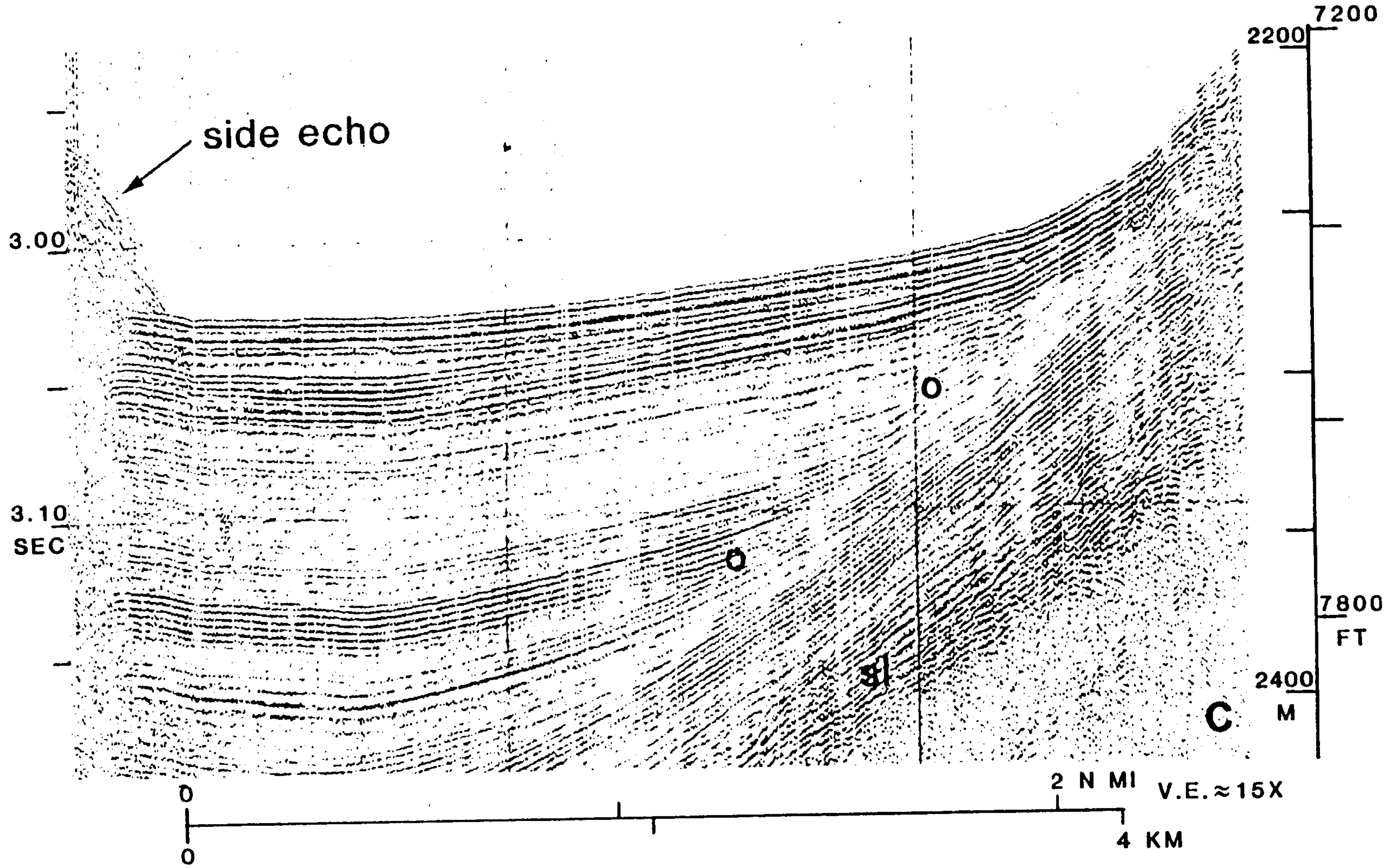


Figure 20. **SEISMIC PROFILE THROUGH A BLOCKED-CANYON INTRASLOPE BASIN SHOWING SLUMPING (s) AND ONLAP OF HEMIPELAGIC SEQUENCE (o)**

After Bouma, Stetling and Feeley, in Bally (1983)

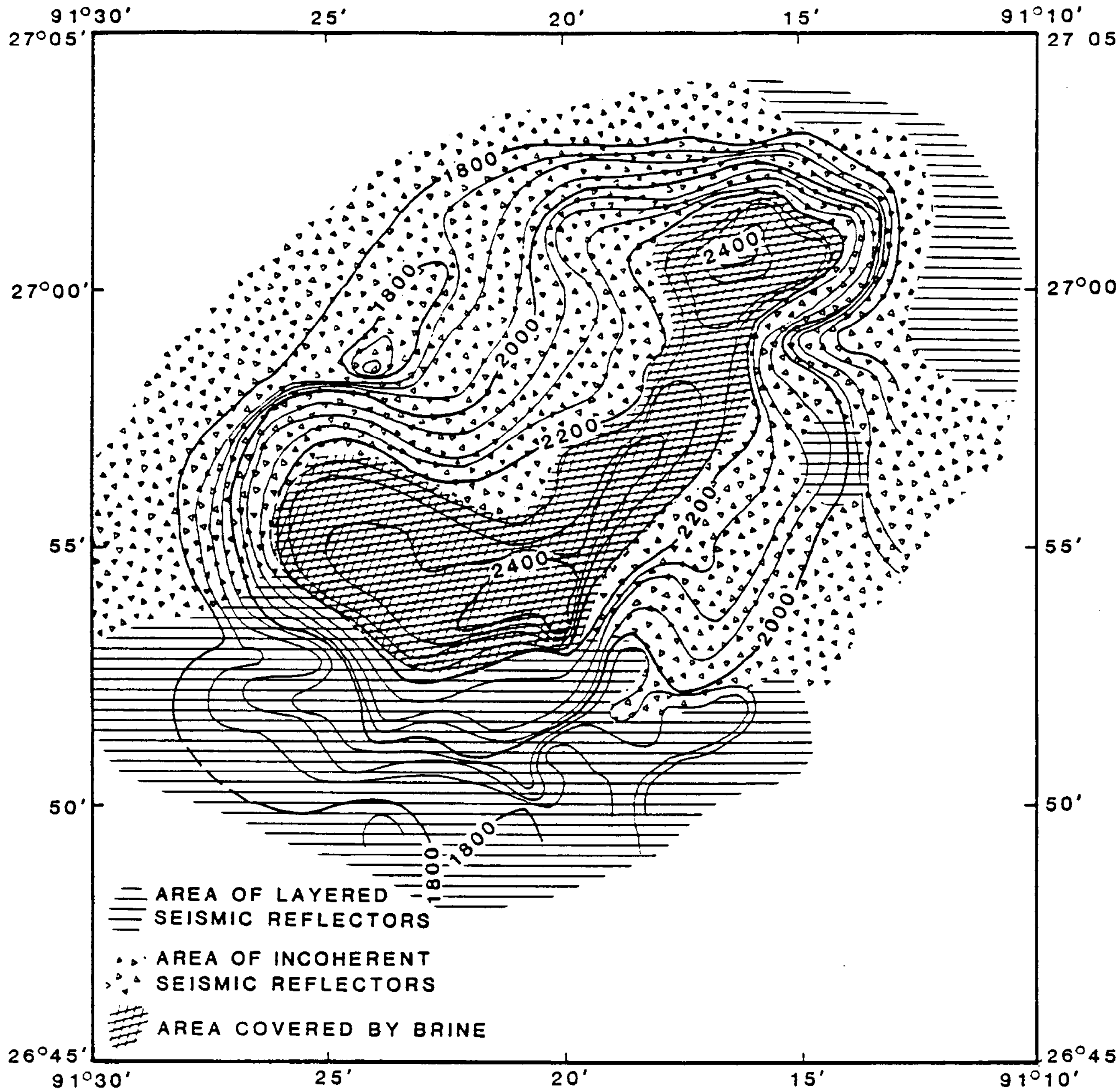


Figure 21. SHALLOW SEISMIC FACIES MAP OF THE ORCA INTRASLOPE BASIN

After Trabant and Presley, 1978

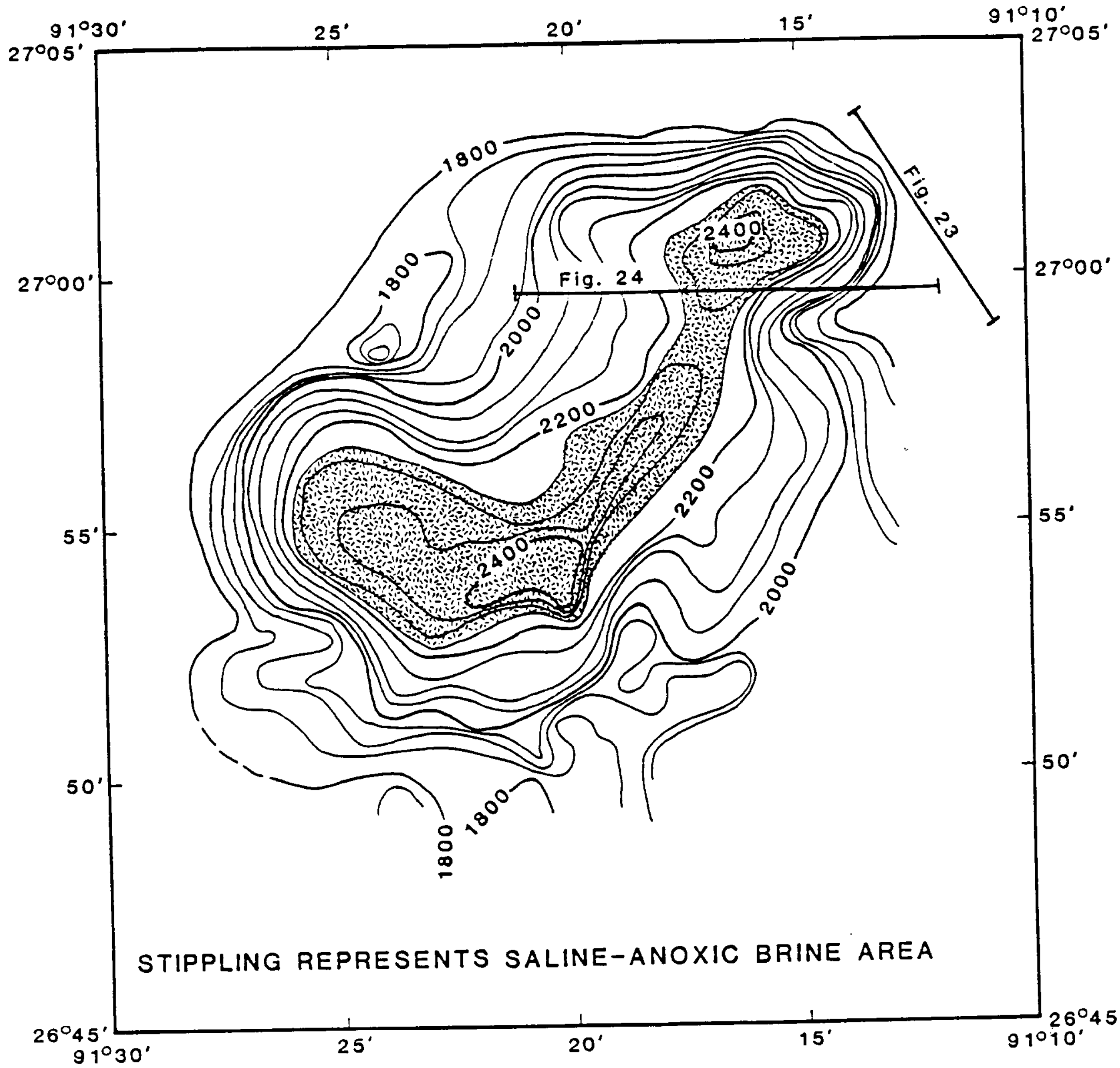


Figure 22. BATHYMETRIC MAP OF THE ORCA INTRASLOPE BASIN SHOWING LOCATION OF SEISMIC PROFILES

After Trabant and Presley, 1978

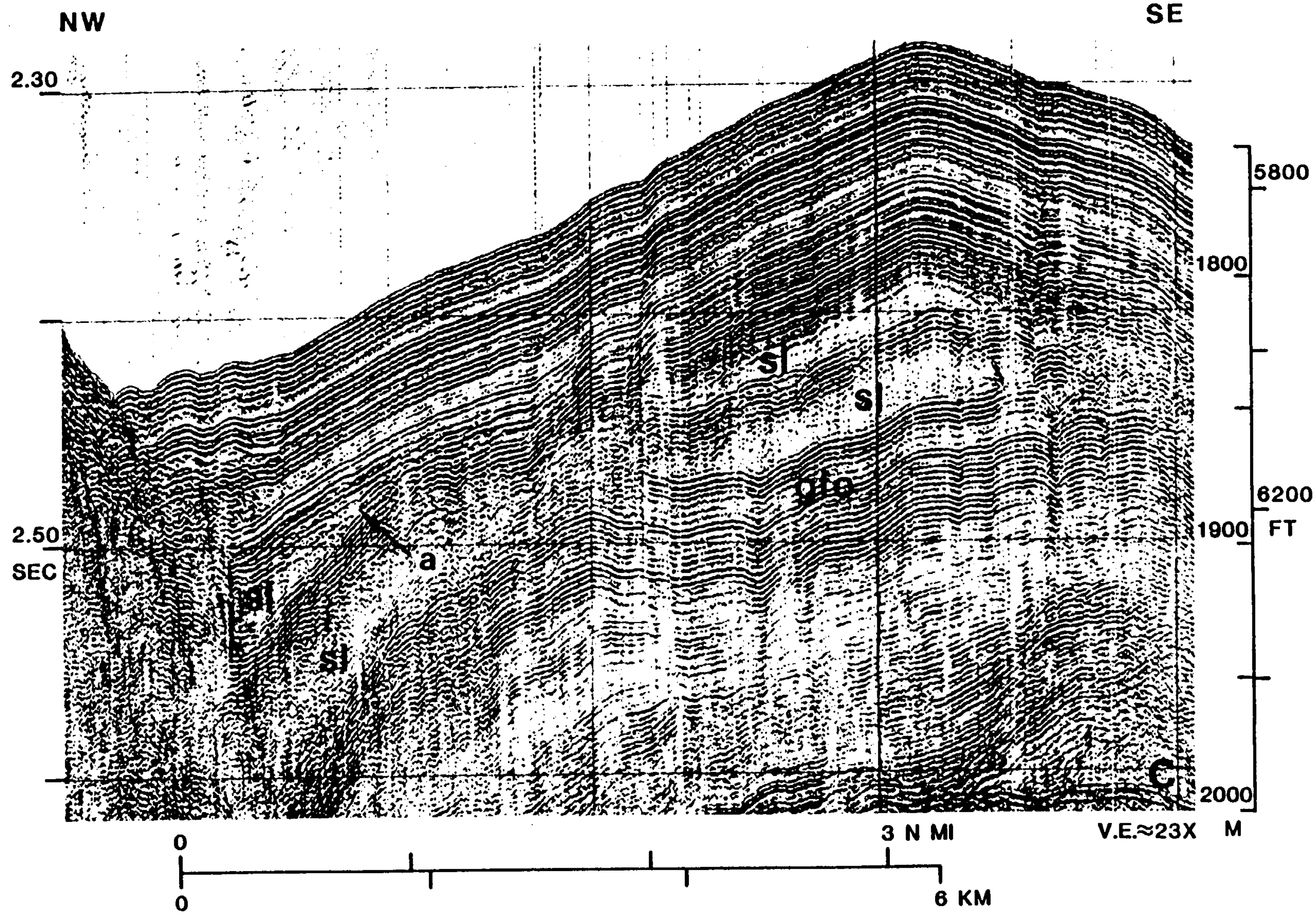


Figure 23. ORCA BASIN SEISMIC PROFILE

After Bouma et al., in Bally (1983)

sl - slump deposit, a - unconformity, gfo - gravity fold

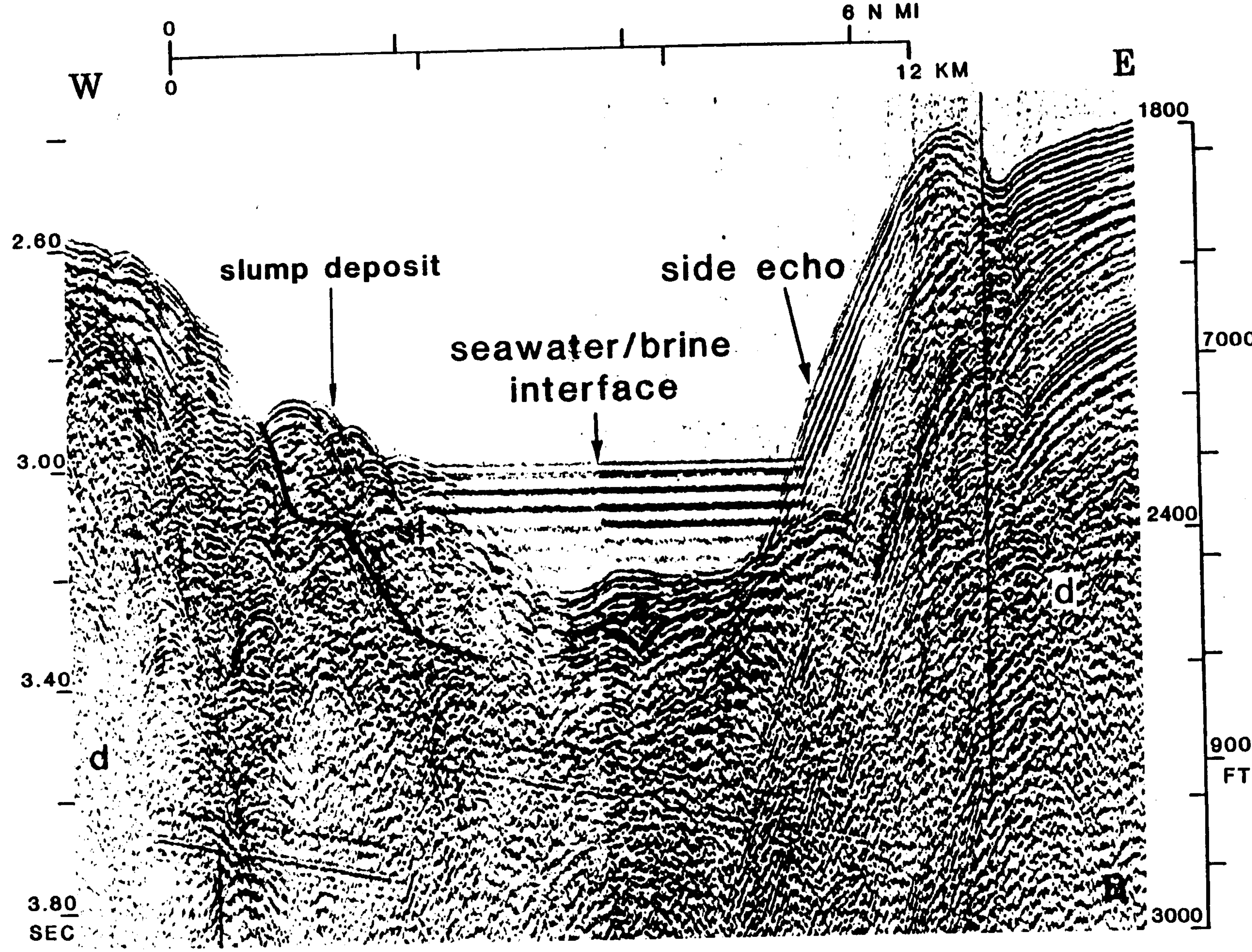


Figure 24. **ORCA BASIN SEISMIC PROFILE**

After Bouma, Stetling and Feeley, in Bally (1983)

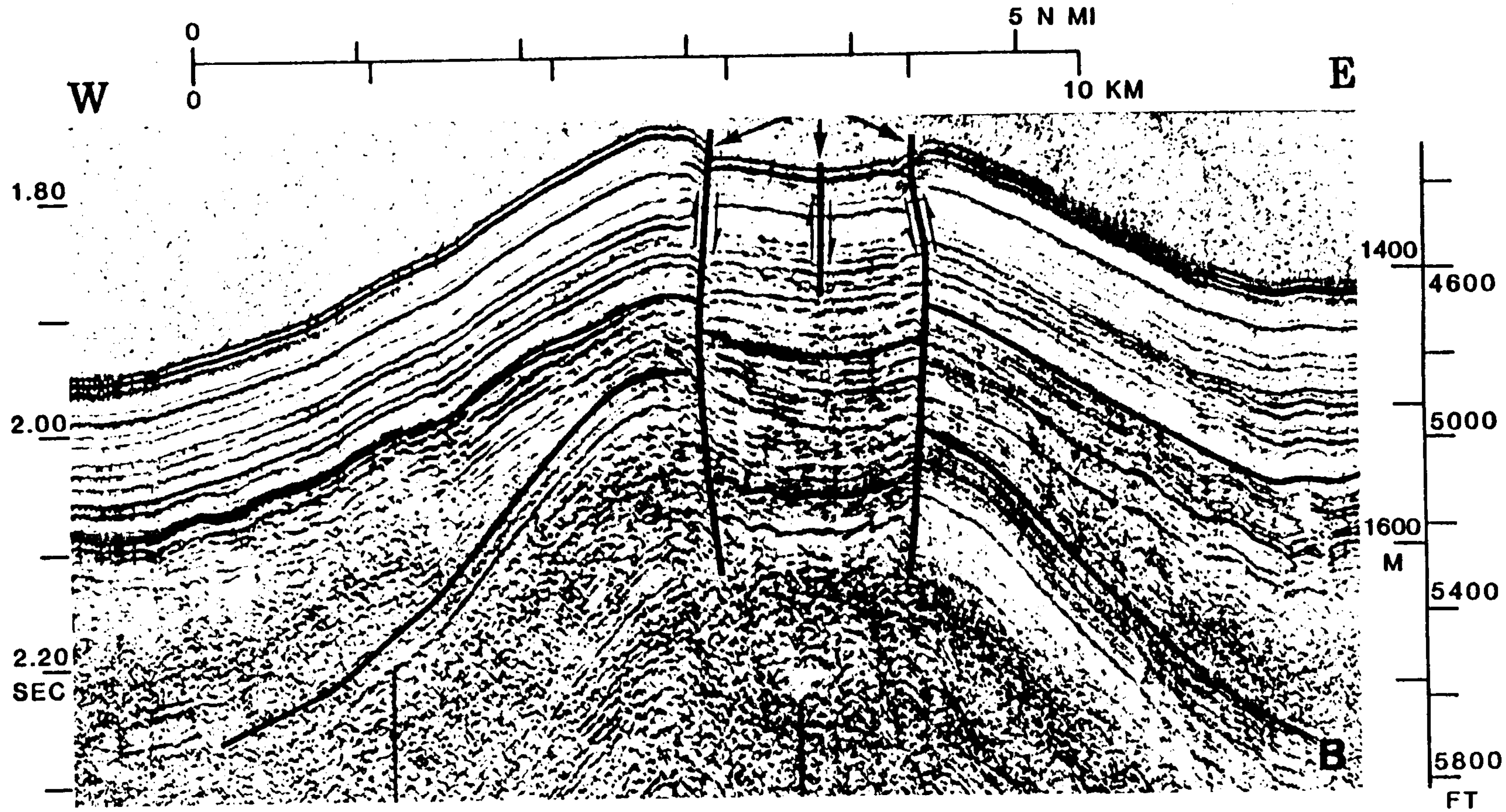


Figure 25. SEISMIC PROFILE THROUGH A SALT DIAPIR COLLAPSE BASIN (CARANCAHUA BASIN)

After Bouma, Stetling and Feeley, in Bally (1983)

discussed below with Pleistocene eustatic sea level fluctuations. The stratigraphic record, as elucidated by high resolution seismic profiling, documents discontinuities, pinch outs, erosional surfaces, chaotic disorganized slump masses, sediment ponding, offlaps, growth faulting, and normal faulting from the basin infill (Figures 20, 23, 24, 26). Slope creep is documented from the flanks of the diapirs which are areas of particular structural complexities (Figure 27). The progressive decrease in deformation from older to younger strata along the diapir flanks, proximal slumping, growth faulting, small-scale thrusts, in association with stratigraphic overlaps, pinch outs, and a tilted angular unconformity document the rise of the salt structures, and the lateral compression exerted by diapir growth (Figures 23, 24, 28). The salt diapirs and the interdiapir basins appear to be constant in a near state of imbalance, sedimentary loading and salt activation continuously interacting with each other.

High resolution seismic-stratigraphic analyses document the detail of these processes, by correlation of various pinch outs, offlaps, and unconformity features. Successive drapes of hemipelagic sediments in different stages of crumbling have been observed indicating late movements. Coring of the sediment section atop some salt domes (Lehner, 1969) has encountered turbidite sands and thick mud sections representing a very high rate of sedimentation which must have been deposited in depression rather than on top of a topographic high. These perched sands and thick mud sequences (150 m in one documented case) support the concept of very rapid and rather erratic growth of salt diapirs through the infill of a former intraslope basin.

The intraslope basin sediment fills are extremely gassy, a condition which enhanced the instability of slope sediments. Seismic wipeout zones (vertical seismic acoustic voids) and sea floor gas mounts are very common features (Figure 29). Brooks and Bryant (1985a) have implied a connection between these seismic features and gas hydrate presence.

Mississippi Fan. The Mississippi Fan is a 170,000 km^2 broad sedimentary apron that transcends neritic, bathyal and abyssal depth zones and dominates the topography of the east central Gulf of Mexico (Martin and Bouma, 1978). This deep sea fan, downslope from the Mississippi River Delta edge is deposited on the continental slope and rise, and on the abyssal plain of the central part of the Gulf of Mexico as the result of the influx of sediments delivered to the Gulf of Mexico by the ancestral Mississippi River since the early Pleistocene.

The continental shelf near the apex of the fan is greatly narrowed by the present Mississippi Delta whose presently active lobe almost entirely crosses the shelf southeast of New Orleans and empties directly onto the continental slope. The apex of the fan is on the uppermost part of the continental slope near the mouth of the Mississippi Trough. From this point, it spreads radially downslope, abutting in its distal portion with the Florida, Campeche and Sigsbee Escarpments, and grading into the near horizontal sea floor of the Sigsbee Abyssal Plain, in 3,500 m water depth (Figure 1). In its median part, the Mississippi Fan has a radium of 350 km, at the 1,200 m isobath, and reaches a maximum thickness of 3,000 m. The topographic slope gradient ranges from 40 m per km on the upper fan to 1 m per km at its outer margin.

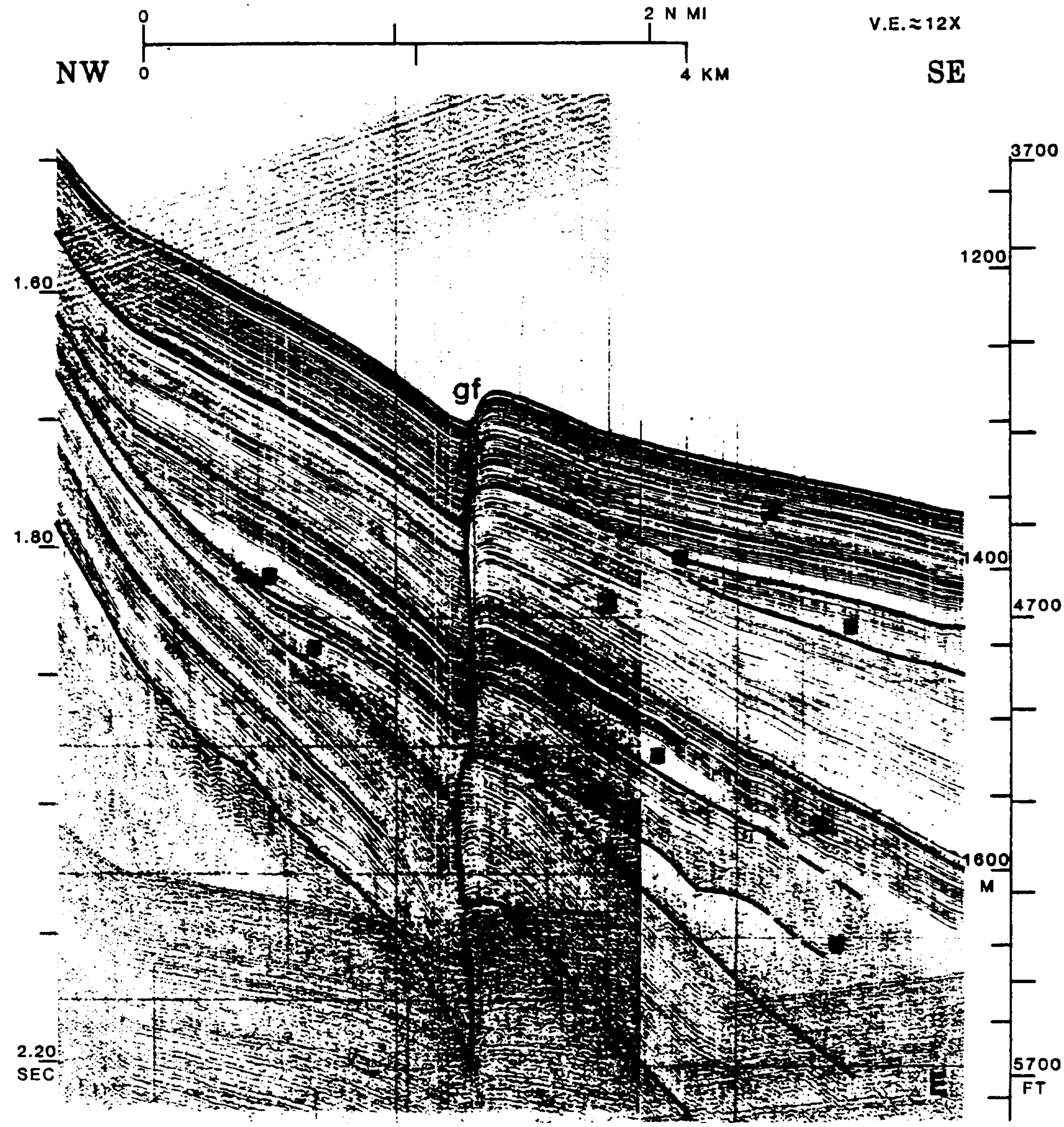


Figure 26. SEISMIC PROFILE SHOWING ANGULAR UNCONFORMITIES AND GROWTH FAULTING (gf)

After Bouma, Stetling and Feeley, in Bally (1983)

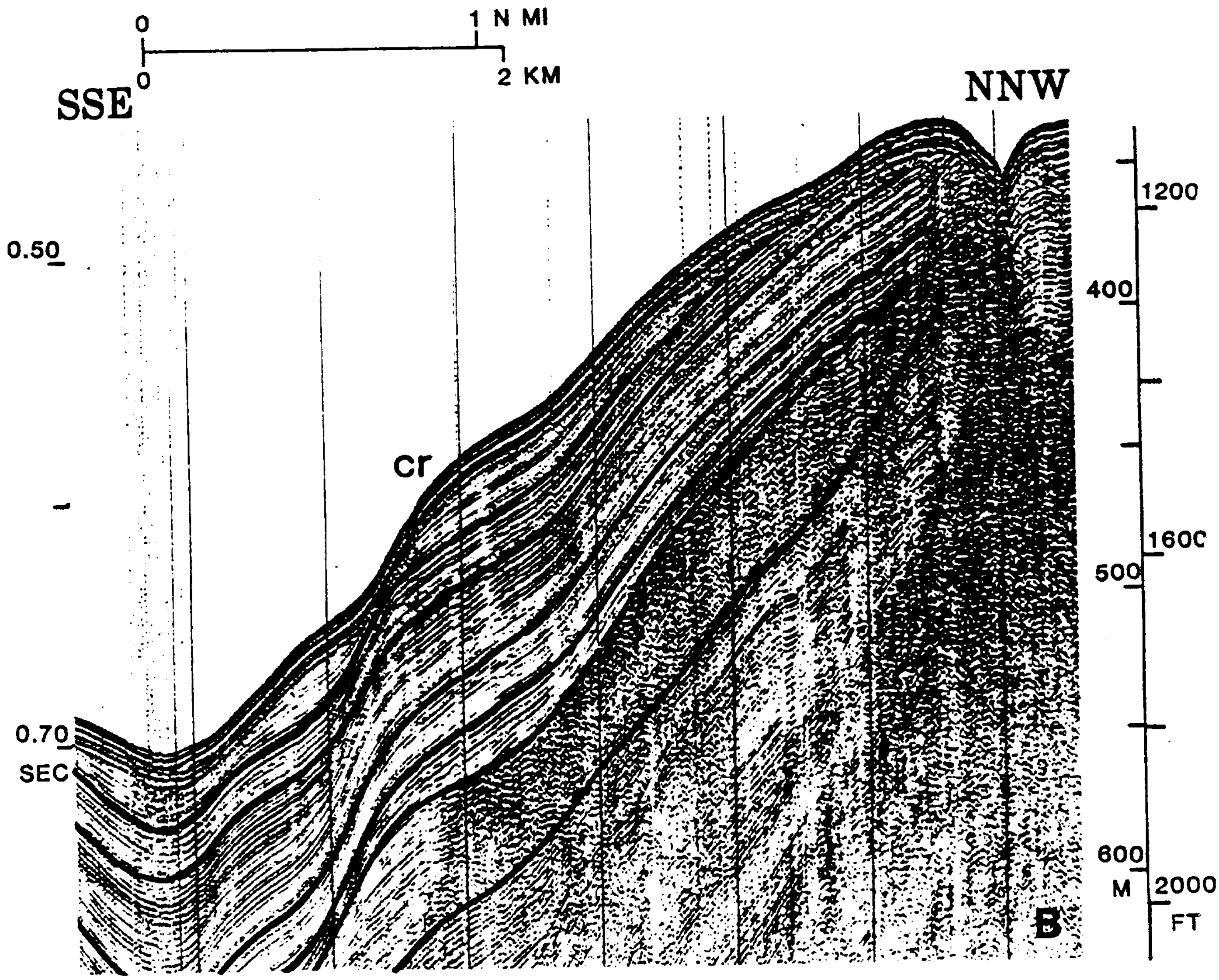


Figure 27. TEXAS CONTINENTAL SLOPE SEISMIC SECTION SHOWING SLOPE CREEP (cr)

After Bouma, Stetling and Feeley, in Bally (1983)

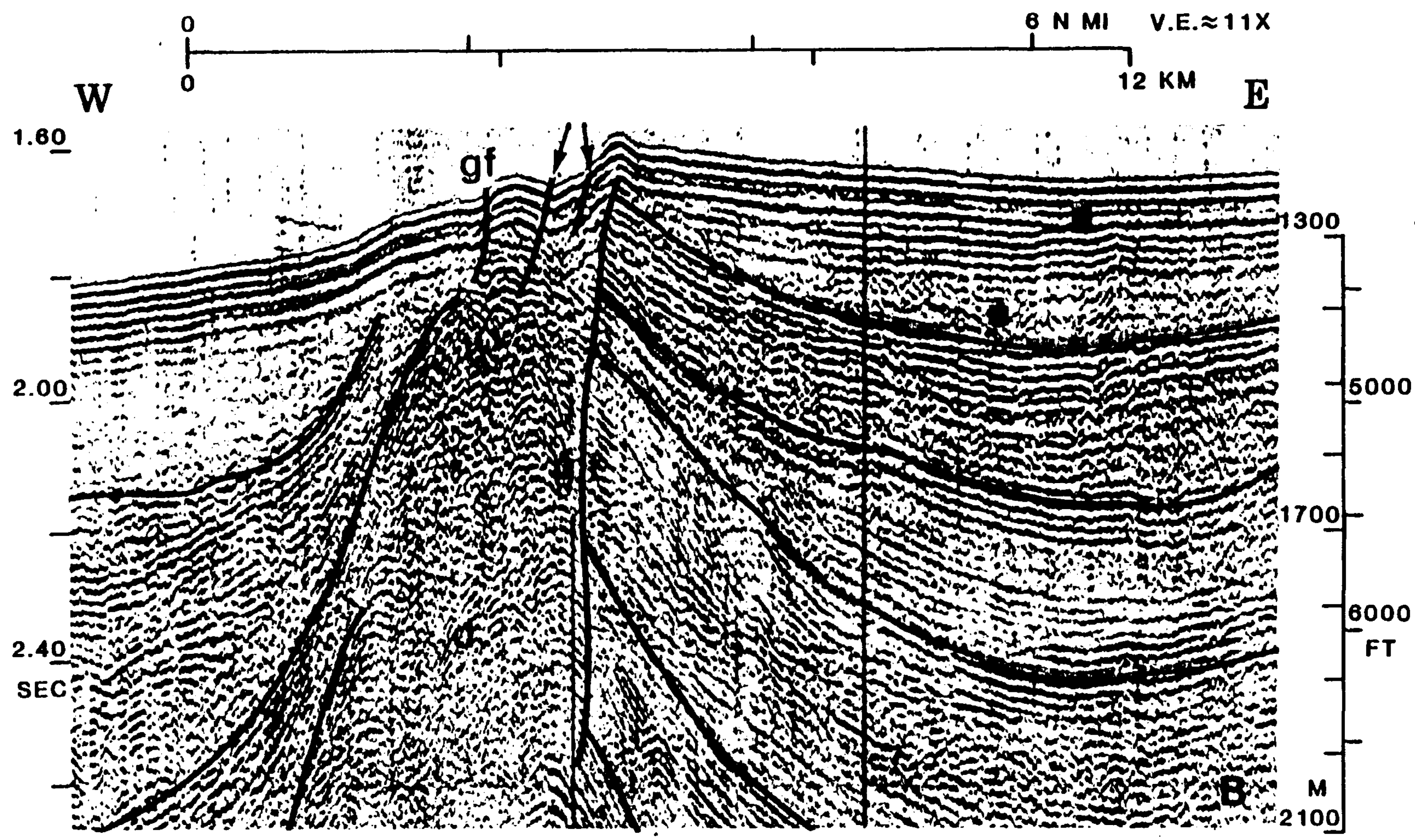


Figure 28. **SEISMIC SECTION THROUGH A DIAPIRIC RIDGE**

After Bouma, Stetling and Feeley, in Bally (1983)

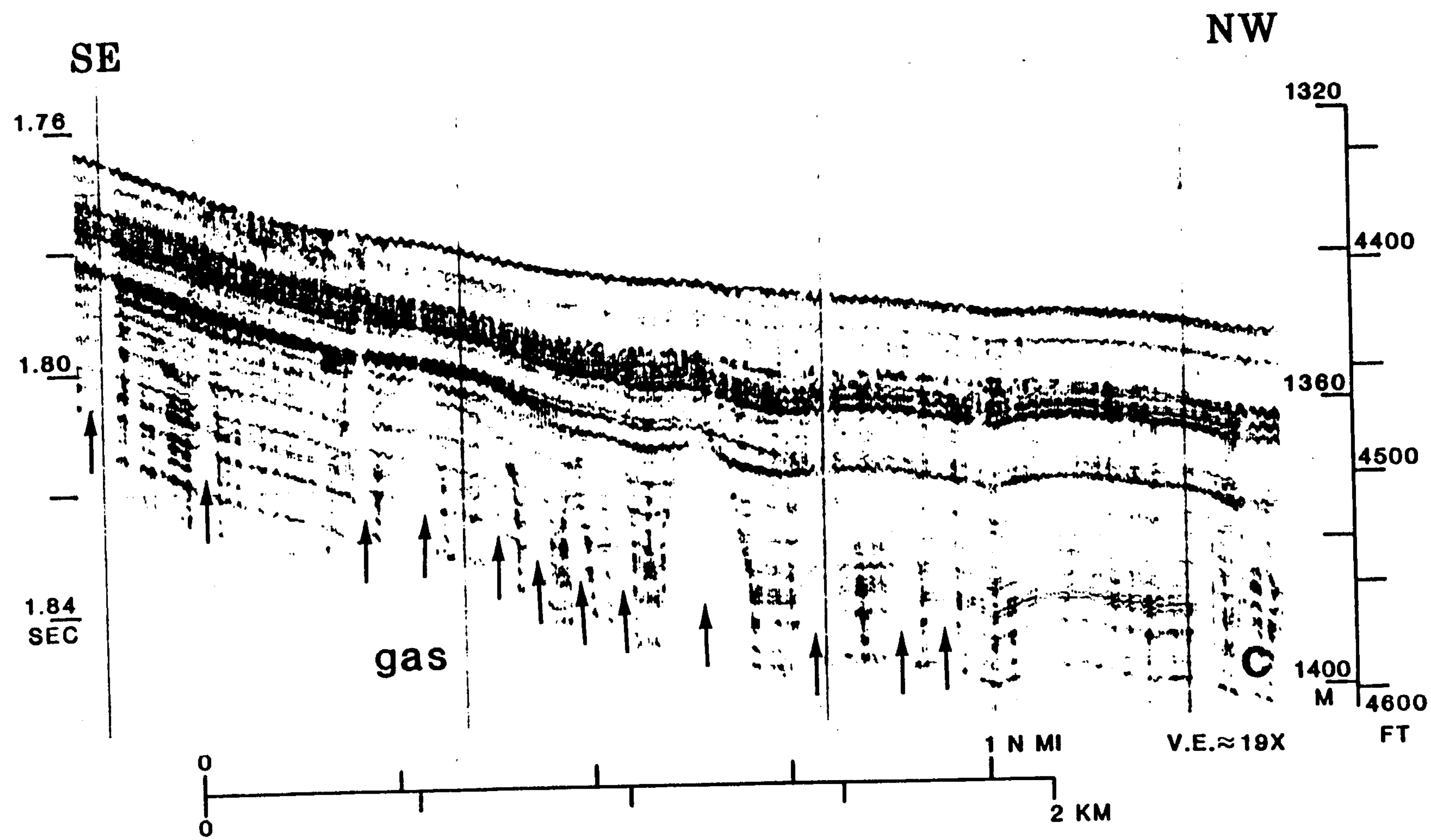


Figure 29. SEISMIC SECTION THROUGH GAS CHARGED SEDIMENTS SHOWING VERTICAL ACOUSTIC WIPE OUT ZONES

After Bouma, Stetling and Feeley, in Bally (1983)

The Mississippi Fan, by virtue of its recent age and large size, heralds a major departure in the geologic evolution of the Gulf of Mexico (Wilhelm and Ewing, 1972, and others). It documents for the first time a general by-passing of the continental shelf and the concomitant cessation of the established shelf edge progradation pattern for this part of the Gulf of Mexico continental margin. More importantly, it represents the first large-scale, direct encroachment of an individualized major clastic system into the abyssal part of the central deep basin of the Gulf of Mexico.

Most of the information presented here on the Mississippi Fan publications of Huang and Goodell (1970), Bouma (1973), Stuart and Caughey (1976), Moore et al. (1978 and 1979), Bouma et al. (1984a and 1984b), and Coleman et al. (1984a and 1984b).

Seismic stratigraphy interpretation of the Mississippi Fan (Moore et al., 1978 and 1979) indicates that the fan is built by a number of elongate fan lobes that are not stacked vertically but switch laterally and prograde basinward. Each of these fan lobes represents a channel-levee-overbank system which can be divided into a canyon and an upper (proximal) fan lobe, a middle fan lobe, and an outer (distal) lower fan lobe. The accretion of the various fan lobe systems led to an overall pattern which is subdivided into three broadly concentric zones labelled upper, middle or supra, and lower fan (Figure 30).

The upper fan is characterized by a large erosive channel cut into older fan sediments, usually nearly filled by late Pleistocene (Wisconsin) sediments and confined between diapiric structures. It grades downslope to a slightly sinuous channel flanked by massive levees. Large-scale recent slumps have affected the upper fan, forming large lobate features and seismically identifiable by a complexly disturbed, chaotic reflector pattern (Walker and Massingill, 1976).

The middle fan represents a large, massive, convex upward contractional complex, formed by the lateral and vertical accretion of individual fan lobes which locally dominate (300 m) the surrounding surface. The hummocky topography probably reflects the irregular surfaces associated with abandoned main and distributary channels as well as by slumping. Individual fan lobe systems are 400 m thick and 150 km wide, and have 3 to 4 km wide, nearly completely filled, sinuous channels at their apexes. The dimensions and sinuosity of the channels decrease downslope. Detailed seismic reflection profiling has documented the meandering of the channels and their associated levees (Kastens and Shor, 1985). Active channels were 100 m deep, and filled at their bottoms by coarse sediment debris flow deposits whose top surfaces show flow line ridges and transverse arcuate ridges where obstacles constrained the flow. These deposits are blanketed by hemipelagic sediments which fill the channel to the brim.

The lower fan system represents an aggradational area where channels and levees become small and shallow and frequently bifurcate before termination. Abandoned channels occur in clusters, paralleling the active channel.

The system of channels briefly outlined above acts as conduit for downslope transport of sand and finer sediments by turbidity currents to the distal portion of the Mississippi Fan. Turbidity currents appear to be constrained to flow within channels but breaching is frequent, especially in the lower fan lobes, and leads to lateral migration. Further downslope, the channel position becomes more instable. Turbidity currents deposit coarse

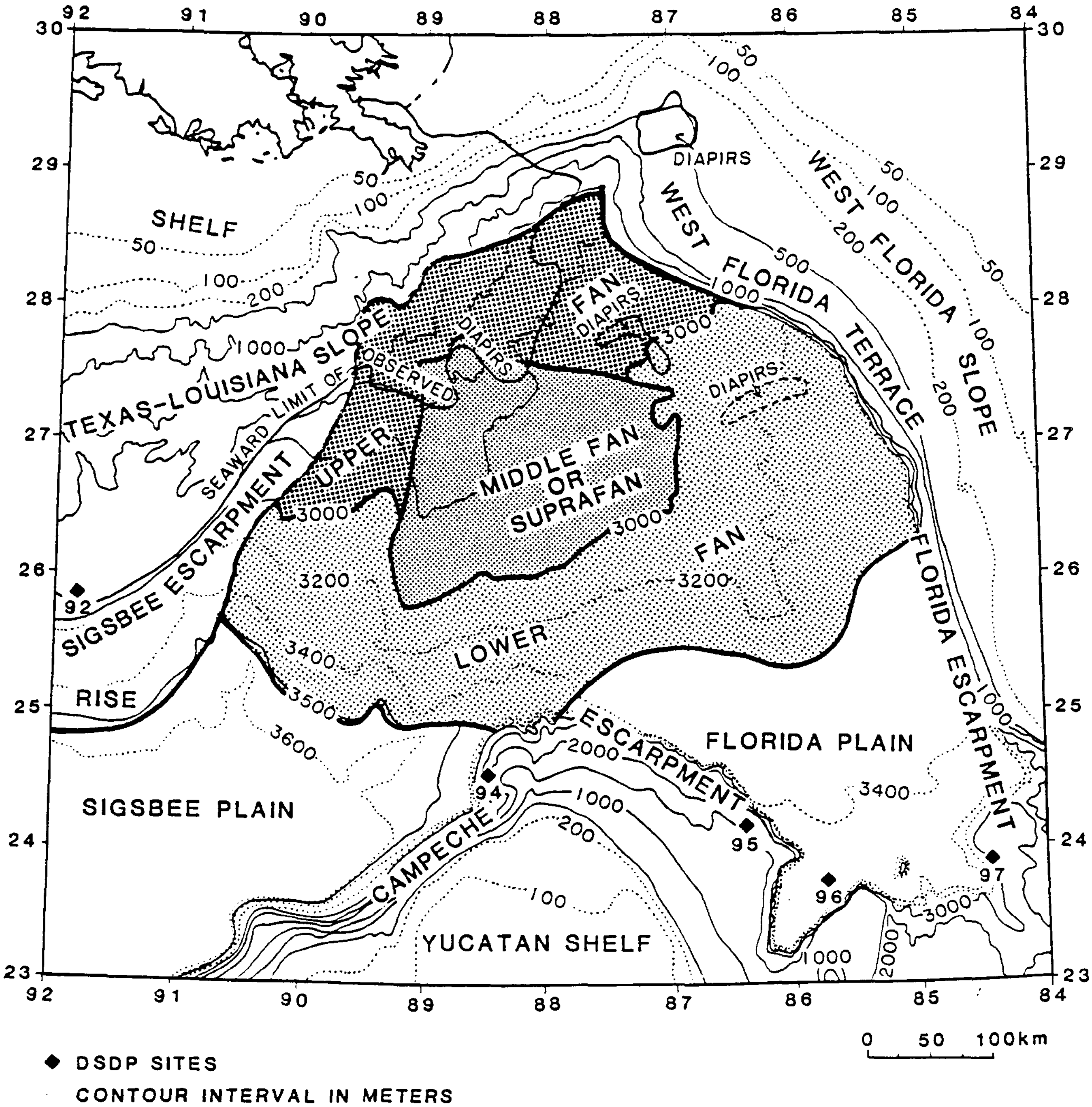


Figure 30. GENERALIZED MAP OF THE MISSISSIPPI FAN

After Moore et al,

sediments in the channel base and finer sediments on the levees. As the volume of supply diminishes the main channel is prevented from any further development as a conduit for sediment transport by clogging by debris flows. The source of clastic material is dominated by clay material and sand, which accumulates within the upper middle fan channel by lateral channel migration and vertical aggradation.

A system of salt ridges representing the buried eastward extension of the Sigsbee Escarpment underlies the Mississippi upper fan. The successive salt ridges diminish in height seaward and pass eastward to isolated salt diapirs, a situation interpreted to reflect the effect of the sediment progradation and load increase on salt migration (Shih et al., 1977).

Mississippi Trough. The apical region of the Mississippi Fan at the level of the present shelf edge is marked in its western part by a conspicuous elongated topographic feature known as the Mississippi Trough or Mississippi Canyon (Figure 1). The Mississippi Trough is a large notch normal to the continental slope. It appears to extend to the 2,000 - 2,500 m isobaths, although it is better defined in its upper course which ranges down to the 1,200 m isobath. It is approximatively 200 km long, 10 to 15 km wide and is characteristically flat-bottomed. This latter characteristic led Ferebee and Bryant (1979) to prefer the designation of trough instead of canyon for this submarine feature. The upper course of the canyon, some 80 to 100 km long has been studied in some detail (Woodbury et al., 1978; Ferebee and Bryant, 1979; Coleman and Prior, 1983; Figure 31).

The Mississippi Trough was originally interpreted as resulting from massive submarine slumping in combination with diapiric positioning (Fisk and McFarlan, 1955). The head of the submarine trough was considered to have been established by subaerial erosion by the ancestral Mississippi River during late Pleistocene (70,000 through 40,000 years b.p.) sea level lowstands. According to this scenario, after the river initially incised the shelf margin, the trough progressively developed by scouring due to slumping and sand-laden density flows generated from the active, instable sedimentary prodeltaic apron of the Mississippi River Delta located at that time near the trough head. Much sediment was funneled through the trough (Wilhelm and Ewing, 1977) and its bathymetric contours were interpreted to reflect the course of the Pleistocene Mississippi River submarine canyon system. Recent investigations by Coleman et al. (1983) have significantly altered this picture. The Mississippi Trough is regarded by these authors as a late Wisconsin erosional submarine feature (post 27,000 years b.p. and pre 20,000 years b.p.) created, in a very short period of time, by retrogressive slope failure and slumping. Successive retrogressive slumping triggered by slope instability led not only to upslope propagation of the trough head but also to downslope partial filling of the newly created depression by slump and debris flow deposits. Once initiated the trough acted as its own conduit, directing the failed material basinward until equilibrium was reached. The trough headwalls reached stability prior to 20,000 years b.p., by which time the latest slump infill, identified by a seismically distinctive topographic surface was completed. The major infill which represents a post slumping depositional phase occured by progradation of a series of late Wisconsin (20,000 through 12,000 years b.p.) submarine delta fan lobes originating north of the trough, from the ancestral Mississippi River Delta. By 10,000 years b.p., the trough was completely inactive and was draped by hemipelagic muds.

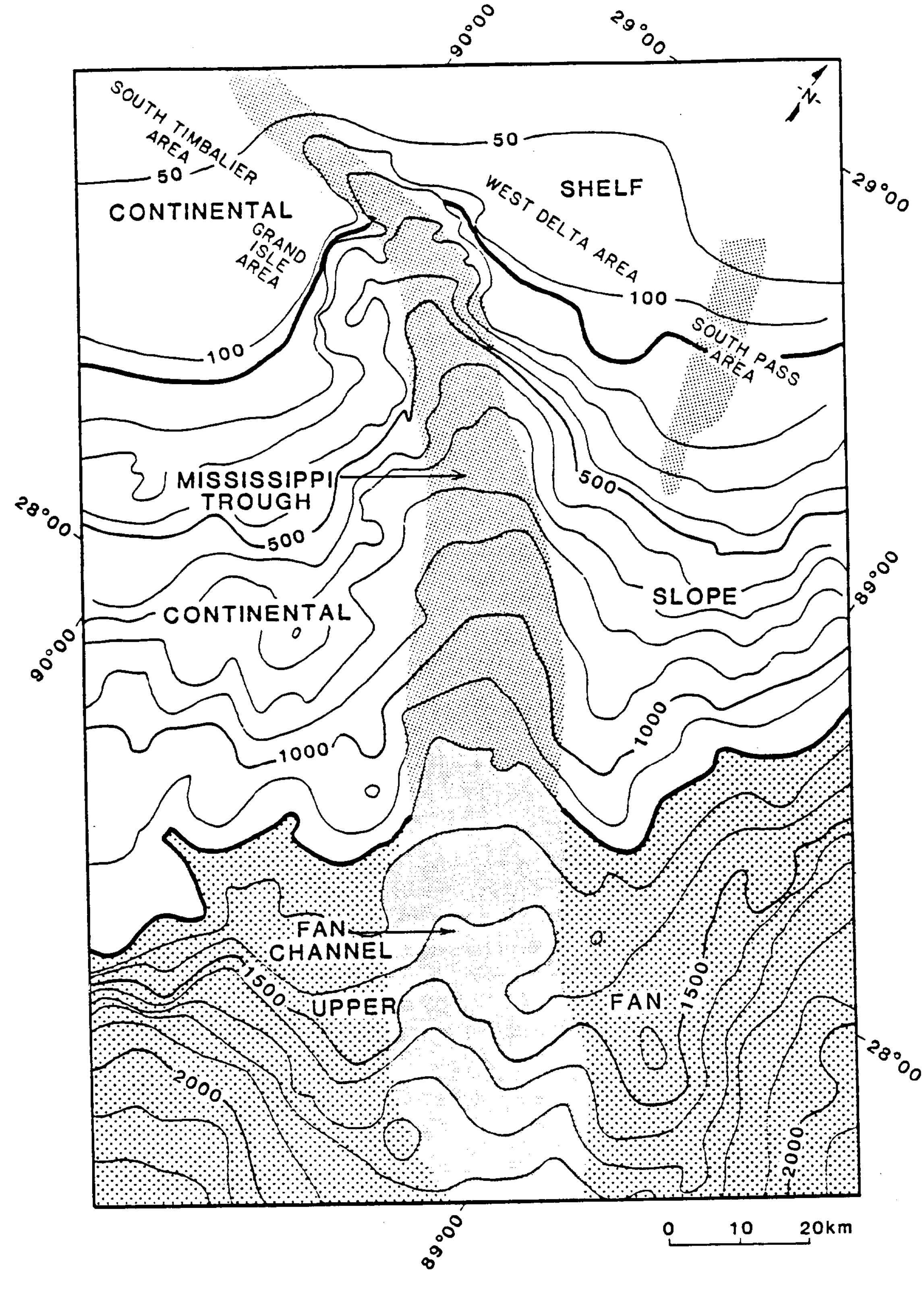


Figure 31. DETAILED BATHYMETRIC MAP OF THE UPPER MISSISSIPPI TROUGH AND ASSOCIATED UPPER FAN CHANNEL

After Woodbury, Spolts and Akers, 1978

The interpretation by Ferebee and Bryant (1979) differs from that proposed by Coleman et al., (1983). Large diapiric masses appear to have had a direct influence on the course and evolution of the Mississippi Trough; small intraslope basins, growth-faults, local slumps on diapir flanks, local sediment ponding, stratigraphic pinch outs against diapiric walls have been documented by Ferebee and Bryant (1979; Figures 32 - 33). The sedimentary infill is very gassy and gas craters, seamounts and very commonly observed seismic voids have also been documented. One acoustically void zone, interbedded within the stratigraphic infill sequence, has been inferred to represent a particularly rich biogenic gas zone. In these geologic and bathymetric conditions, the occurrence of gas hydrates has to be expected. Gas hydrates have been recovered from the Mississippi Canyon lease area in 1,345 m of water (Brooks and Bryant, 1985a). The exact location of the hydrate recovery remains proprietary.

The Western Margin of the Gulf of Mexico

The western margin of the Gulf of Mexico is bordered inland by the Sierra Madre Oriental, a Late Cretaceous orogenic belt of folded and thrusted Middle Cretaceous Platform limestones overlapped to the east by Cenozoic clastic deposits. The Cenozoic clastic apron underlies the continental shelf and continental slope, gently merging with the Sigsbee Abyssal Plain and Gulf of Campeche. The eastern Mexico continental slope is characterized by a unique geologic feature, the Mexican Ridges. The Mexican Ridges form an enormous belt of folded Tertiary strata that extends along the entire length of the western Gulf of Mexico continental slope from about 25°N latitude to about 19°N latitude (Figures 1 - 2). The folds form long, linear, subparallel topographic features on the sea floor. The nature and origin of these folds has been the subject of debate since their discovery in the 1960s.

Geologic Evolution

The very thick deformed Cretaceous limestone strata of the Sierra Madre Oriental represent the western portion of the Cretaceous carbonate platforms which rimmed Gulf of Mexico (Coogan et al., 1972; Enos, 1974; Enos, 1981; Viniegra and Castillo-Tejero, 1970; Viniegra, 1981). Individual carbonate platforms, with their restricted marine shelves and evaporitic lagoons, such as the Valles, San Luis Potosi platforms, and the Golden Lane Atoll, were flanked seaward by steep constructional edges with talus deposits at their toes (Figure 2).

The shallow water carbonate platforms were drowned during the Turonian and covered by pelagic limestones which graded upward to clastic sediments. By the early Tertiary the clastic sediments began to encroach over the Gulf of Mexico. Marginal basins developed in front of the Sierra Madre Oriental, (Mossman and Viniegra, 1976; Helu et al., 1977; Busch and Govela, 1978). As seismic stratigraphic studies indicate (Buffler, 1983; Buffler et al., 1979), some material was redistributed and deposited offshore of these marginal basins within what is presently the eastern Mexico continental shelf and slope and the deep central basin. By late Miocene time, the clastic influx spilled over

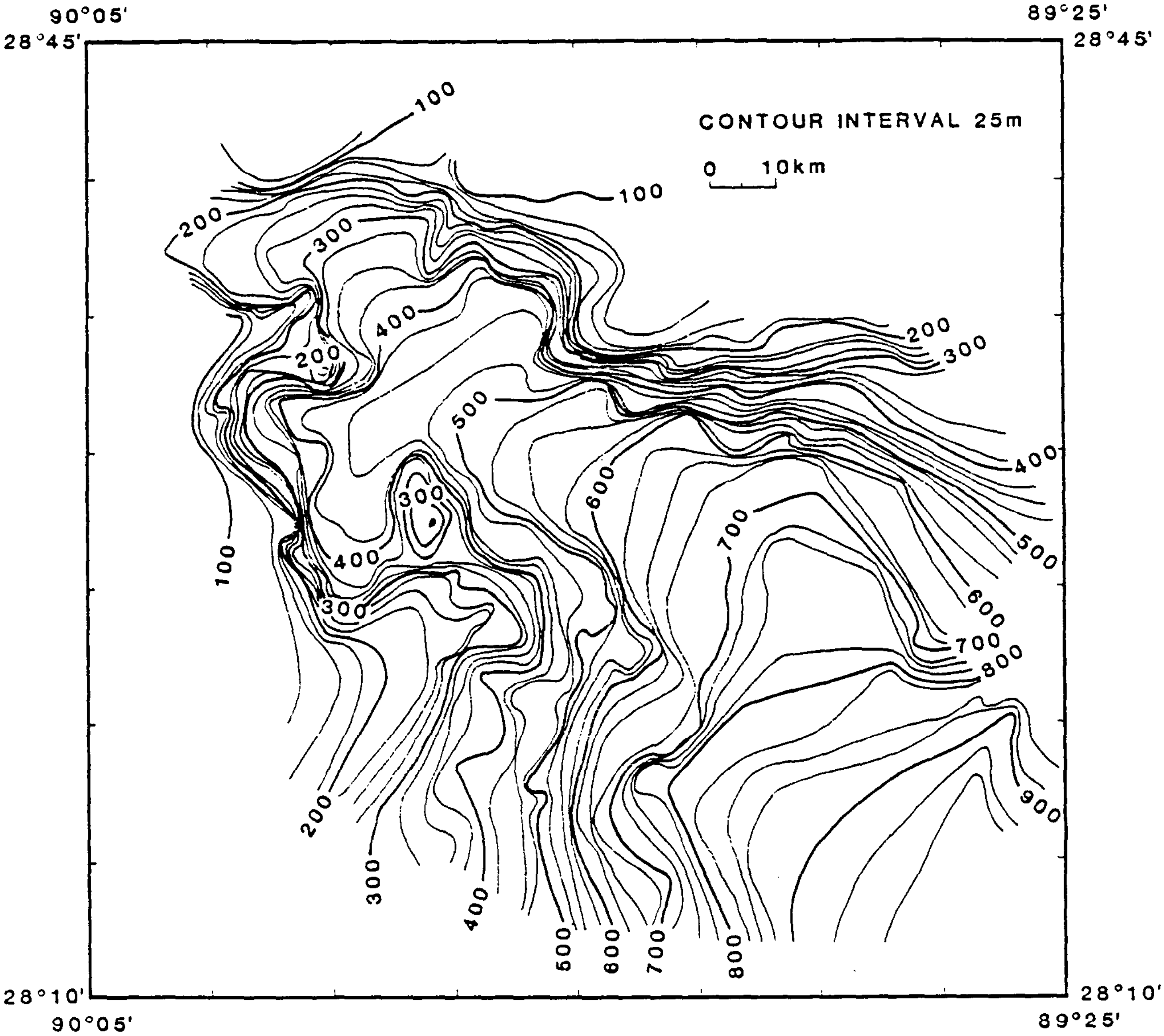
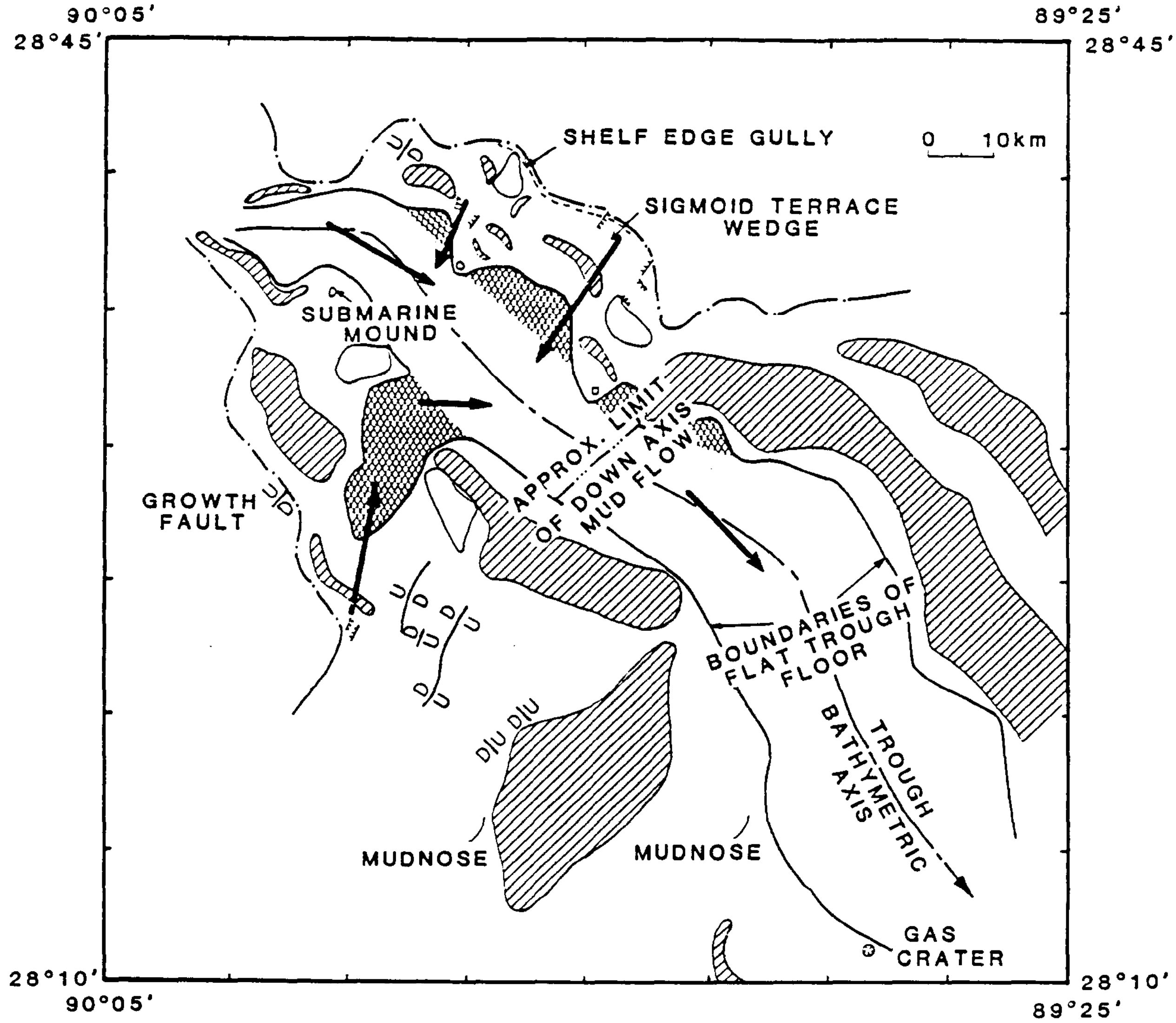


Figure 32. **DETAILED BATHYMETRIC MAP OF THE UPPER MISSISSIPPI TROUGH**

After Ferebee and Bryant, 1979



SHELF MARGIN AND EROSIONAL ESCARPMENT

MASS MOVEMENTS OF SEDIMENTS SLUMPED AND FAULTED

SEMI-ENCLOSED BASINS AT BASE OF WALLS

SEDIMENT TRANSPORT DIRECTIONS

FAULTS WITH RELATIVE MOVEMENT UP-DOWN

DIAPIRS

SMALL MUD DIAPIRS

BOUNDARIES OF EROSIONAL CHANNELS

Figure 33. **PHYSIOGRAPHIC AND GEOLOGIC FEATURES OF THE MISSISSIPPI TROUGH**

After Ferebee and Bryant, 1979

the still subsiding marginal basins, and the entire margin prograded over the deep Gulf of Mexico. Seismic sections through the Mexican Ridges show very regular reflectors in the upper part of the section from the middle Tertiary onward. This sequence has been interpreted as an alternation of sandy and fine-grained distal turbidites and hemipelagic silty clays, merging with seismic continuity into the section underlying the Sigsbee Abyssal Plain (Figure 34). The isobath maps of the deep Gulf seismic unit by Shaub (1984; Figures 9 through 13) document the existence of an active depocenter in the southwest part of the Gulf of Mexico during middle Tertiary time.

Mexican Ridges

The Mexican Ridges were discovered in the mid-1960s and have been the subject of interest since (Jones et al., 1967; Bryant et al., 1968; Massingil et al., 1973; Hey, 1975; Buffler et al., 1979; Pew, 1982; Buffler, 1983).

The role of salt tectonics in the formation of the Mexican Ridges and the presence or absence of salt under the eastern Mexico continental slope have been debated by many authors: Antoine and Bryant (1969), Antoine and Pyle (1970), Antoine (1972), Bergantino (1971), Ensminger and Matthews (1972), Wilhelm and Ewing (1972), Garrison and Martin (1973); Moore and Castillo (1974), Watkins et al., (1975), Watkins et al., (1976, 1977 and 1978), Martin and Bouma (1978), Worzel and Burk (1979), Winker and Edwards (1983), Shaub (1983), Bertagne (1984).

Detailed seismic profiling of the Mexican Ridges was published by Buffler et al. (1979). Seismic bottom simulating reflectors in some of the anticlinal ridges were identified as possible gas hydrate reflectors. An unlocated BSR, possibly from the area, was also published by Hedberg (1980; Figure 35).

Physiography and Geology. The Mexican Ridges are characterized by gentle folding of late Cenozoic sediments. The folds, linear to slightly arcuate in the southern part of the fold system, have topographic expression up to 700 m, and are extraordinarily regular (Figures 36 - 37). The entire length of the fold system approaches almost 500 km and the average width varies between 50 to 75 km. However, the total extent of the fold system is not precisely known; the folds disappear under recent sediments at the northern and southern end of the system. The fold belt parallels the shelf and dominates the middle portion of the eastern Mexican continental slope. The folds have wave lengths of approximatively 10 km and are quite symmetrical, but in some traverses show slight asymmetry in both a landward and a seaward direction, with the steeper flanks seaward being the most common configuration. Some folds are bounded seaward by distinct thrust faults (Figure 36). The faults dip landward and tend to flatten at depth. Most of these faults cut up through the cores of anticlines. They are very steep in the axes of the tighter folds and are much less steep in the seaward gentler folds. The recent investigation by Buffler et al. (1979) from the southern Mexican Ridges indicated that the folds in this region are less regular than previously believed and change in form along their strike. Also, the wave length and amplitude are less constant than previously reported. The higher amplitude folds begin abruptly midway down the slope, decrease in amplitude seaward, and finally die out at the level of the lower continental slope.

The strata affected by the folding apparently consist of an alternation of sandy and fine-grained turbidites and hemipelagic sediments. The age of this

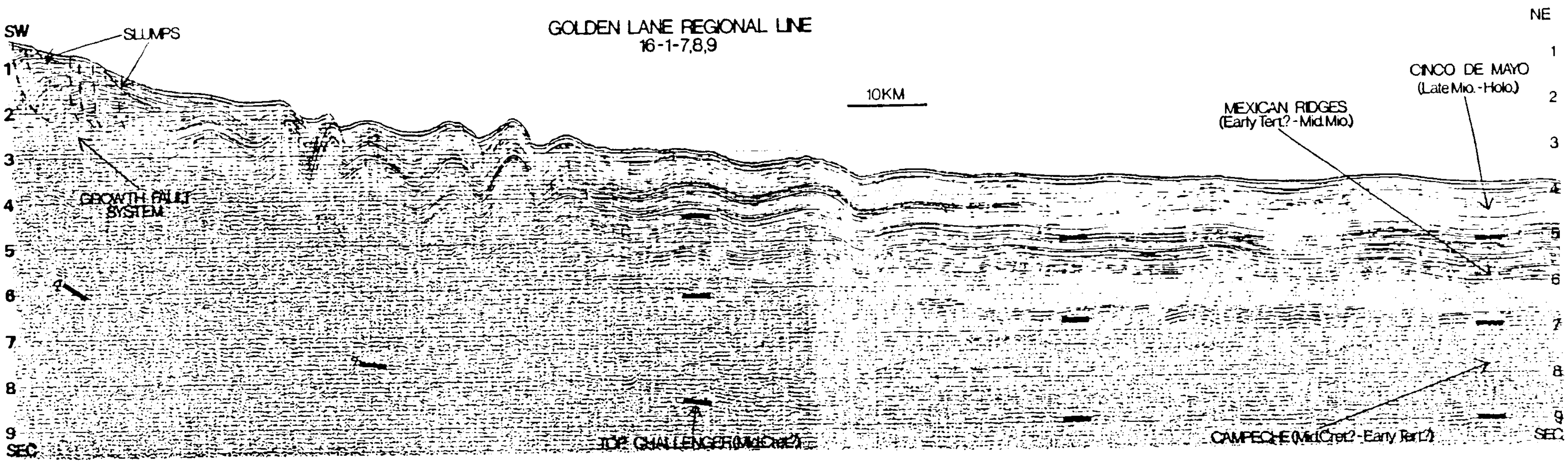


Figure 34. SEISMIC PROFILE SHOWING THE STRUCTURE OF THE MEXICAN RIDGES FOLDBELT, SOUTHWEST GULF OF MEXICO

After Buffler, in Bally (1983)

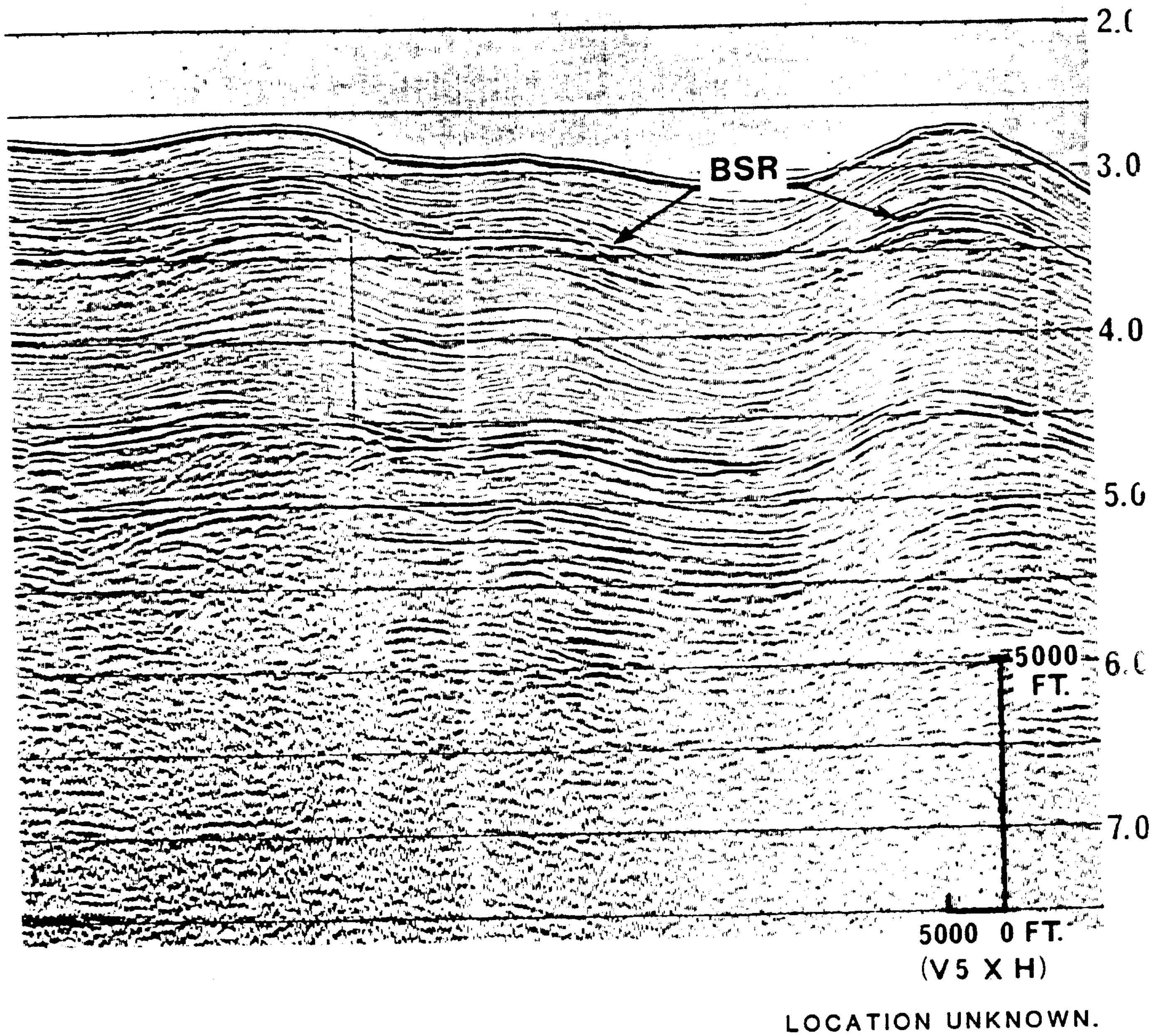


Figure 35. SEISMIC PROFILE SHOWING A BOTTOM SIMULATING REFLECTOR

After Hedberg (1980)

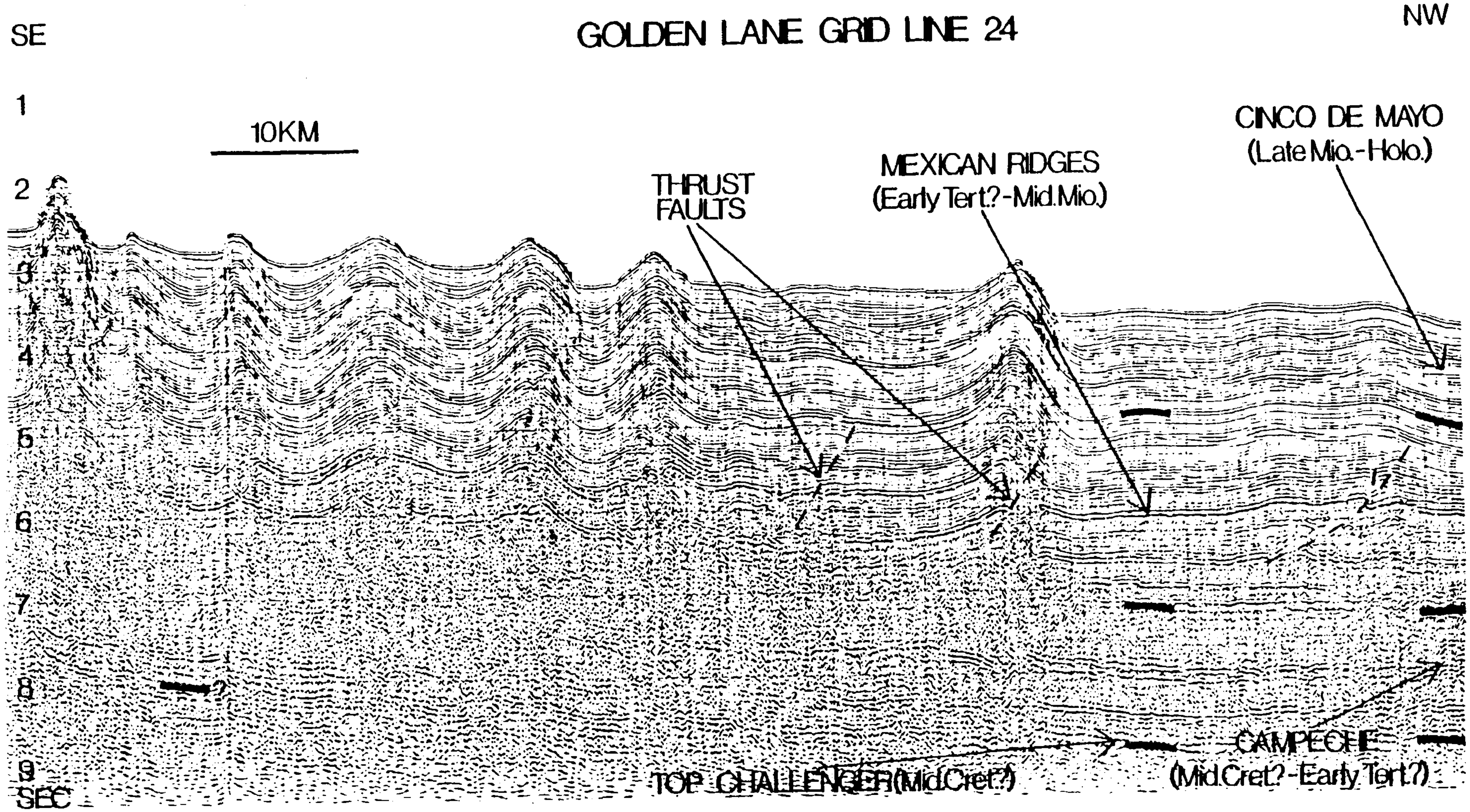


Figure 36. SEISMIC PROFILE SHOWING THE STRUCTURE OF THE MEXICAN RIDGES FOLDBELT, SOUTHWEST GULF OF MEXICO

After Buffler, in Bally (1983)

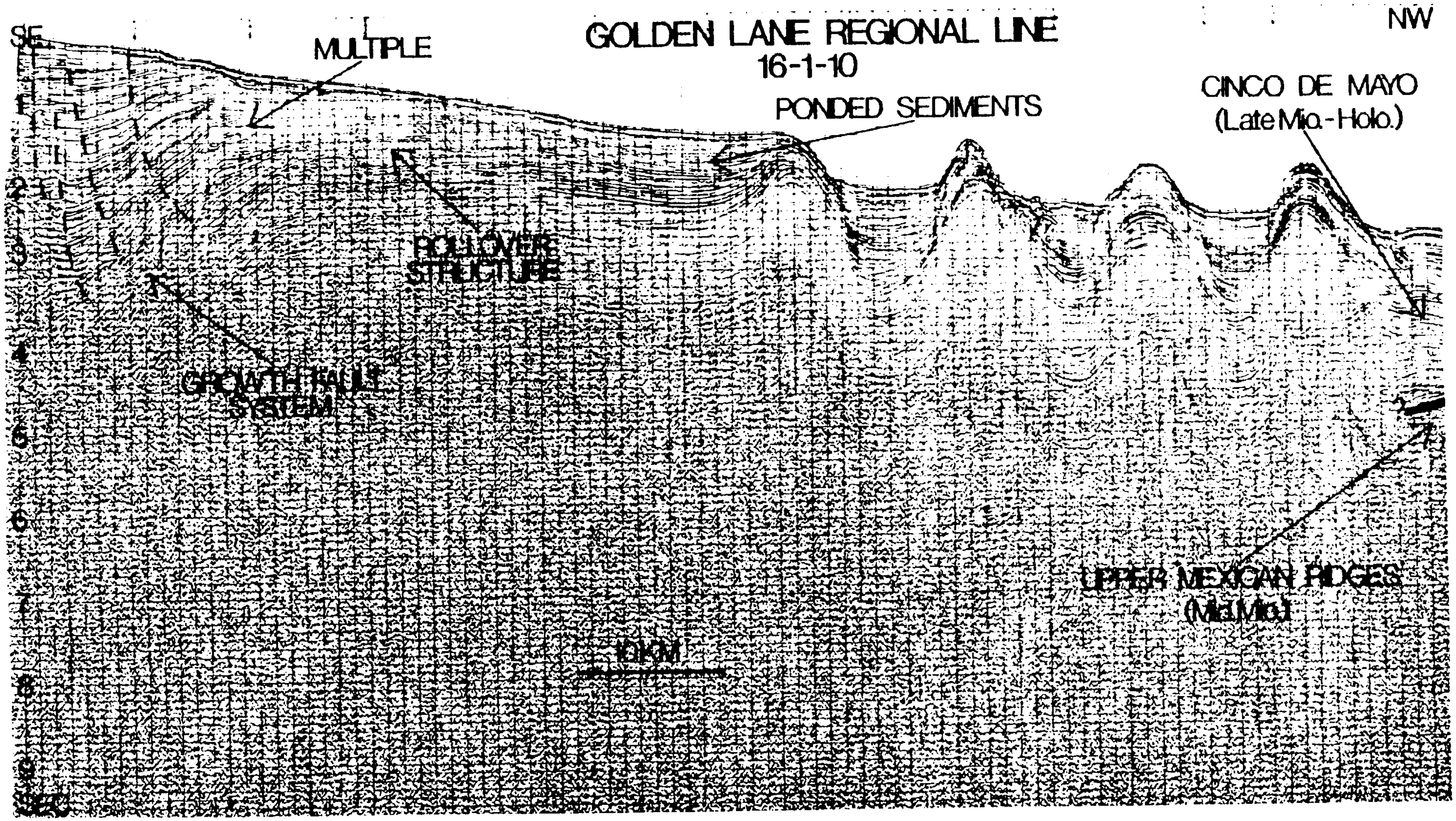


Figure 37. SEISMIC PROFILE SHOWING THE STRUCTURE OF THE MEXICAN RIDGES FOLDBELT, SOUTHWEST GULF OF MEXICO

After Buffler, in Bally (1983)

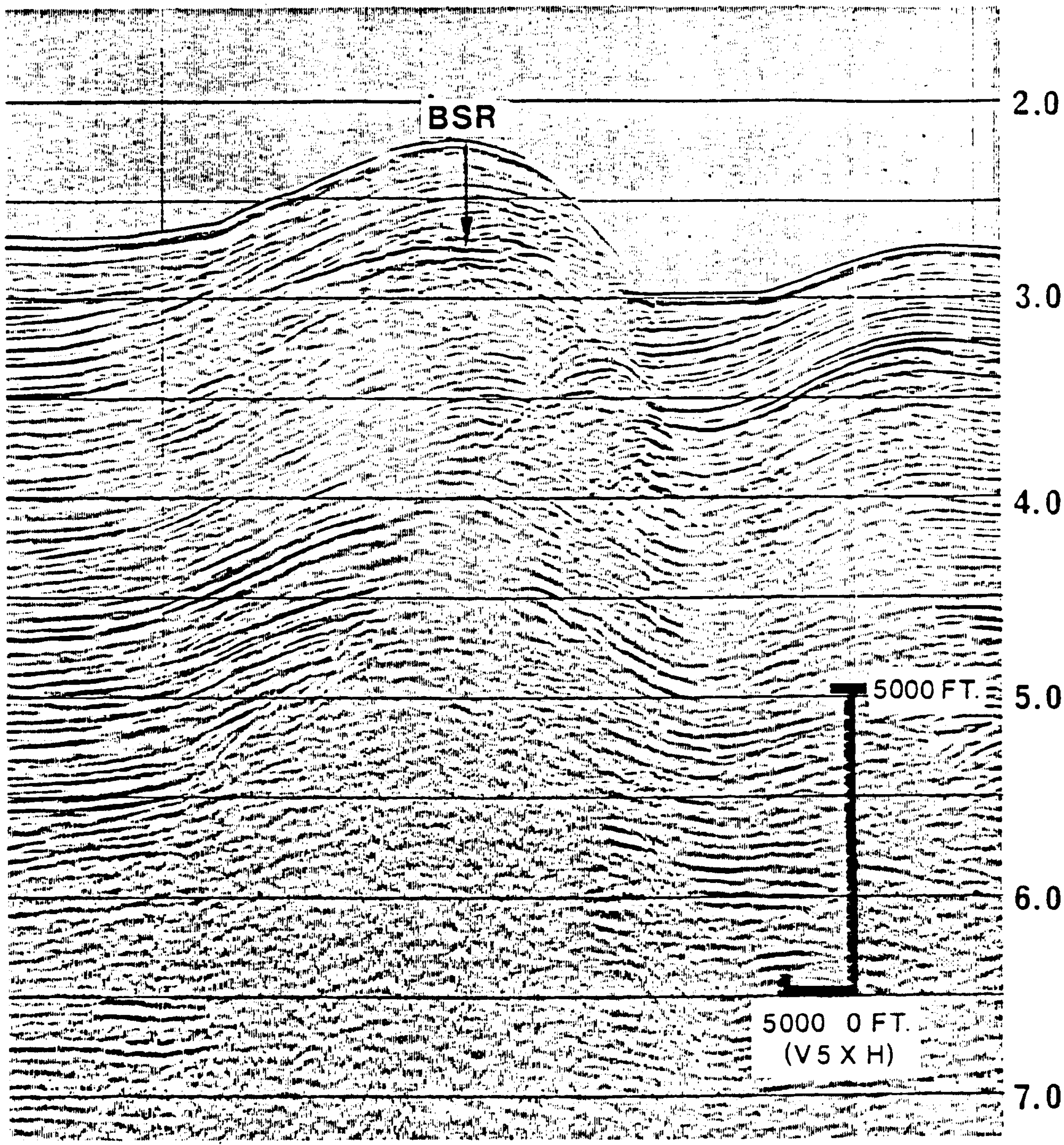
extend under the outer part of the Mexican Ridges, and some deep reflectors under the central fold belt exhibited seismic characteristics typical of the Challenger (Buffler et al., 1979, among others). These authors stated that the top of the Challenger unit continues independently and relatively undeformed beneath the folds, with one noticeable exception.

Regional seismic profiling by Watkins et al. (1977) from the Tamaulipas outer continental shelf and continental slope (Mexican Ridges) through the deep Gulf of Mexico Basin ending at the Campeche Escarpment clearly showed seismic structures under the Tamaulipas upper slope similar to the seismic expression of the diapiric salt masses of the Texas - Louisiana continental slope. Salt diapirism under the Tamaulipas upper slope was recognized by Winker and Edwards (1983) and seriously considered as an alternative to shale diapirism by Shaub (1983). The possibility of the salt unit extending farther offshore under the Mexican Ridges appears likely, in light of the presence of a deep salt-bearing unit (Challenger) within the central Gulf basin.

The evidence for the decollement hypothesis based on fold configuration was cited by Buffler et al. (1979) and Buffler (1983).

The decrease in fold amplitude and dip of the imbricate thrusts in a seaward direction suggests that the tighter folds updip formed first and the gentler folds downdip formed later (Figure 36). This is supported by the ponding of younger sediments updip, while equivalent beds appear to be gently folded along with the entire section farther downdip. The geometry of the Mexican Ridges foldbelt suggests that the deformation has been caused by compressional stresses acting in an east-west direction with large-scale instability of slope sediments causing massive gravity sliding along geopressurized shales (Buffler et al., 1979; Buffler, 1983) or a salt layer.

The evidence for mudstone mobilization is the presence of a few sea floor mud volcanoes in the southernmost part of the Mexican Ridges, offshore of Veracruz (Moore and Castillo; 1974). One unlocated sea floor mud volcano ('shale diapir'), with a BSR has been reported by Hedberg (1980), possibly from this area (Figure 38).



LOCATION UNKNOWN.

Figure 38. SEISMIC PROFILE THROUGH A MUD DIAPIR

After Hedberg (1980)

PART II

FORMATION AND STABILITY OF GAS HYDRATES

Occurrence of Thermogenic Gas Hydrates

The Gulf of Mexico is the only region where gas hydrates composed of thermogenic natural gas have been proven to exist. Workers from Texas A&M have recently recovered sediment cores containing thermogenic hydrates from three sites in the Gulf continental slope approximately 200 km south of the Louisiana coast in 560 to 864 m of water (Brooks and Bryant, 1985a). The recovered gas hydrate samples were associated with biodegraded oil stained sediments. The hydrates ranged in size from disseminated grains to nodular chunks several centimeters in diameter.

Other possible occurrences of gas hydrates can be inferred from the results of DSDP Leg 10. Substantial amounts of gas were encountered while drilling four of the 13 holes of Leg 10: at Sites 88 on the Campeche Knolls, Site 89 in the Gulf of Campeche, and Sites 90 and 91 in the Sigsbee Abyssal Plain (Worzel and Bryant, 1973). When cores from these holes were brought on board ship, anomalous degassing characteristics were observed. The cores emitted very large quantities of gas requiring special measures to prevent complete disruption of the core sediments. The degassing of these cores took place over an abnormally long period of over two hours. The shipboard scientists concluded that it was likely that the cores contained hydrates: "At the time of core recovery, these results were perplexing, but in retrospect it now seems probable that gas hydrates were present in these cores" (Worzel and Bryant, 1973). Project scientists concluded that all gassy cores contained microbially produced gas based on the vertical distribution of gas in the sediments (Beall et al., 1973). The gas content of the sediments reached a maximum at 200 to 400 m depth then diminished with depth. With no overlying lithologic seal, the authors assumed that thermogenic gas migrating from deep sources would be more concentrated near the source. This would have resulted in a trend toward greater concentration of gas with depth, unlike that which was observed. In situ generation of methane by bacteria would produce a maximum of interstitial gas where growth conditions were optimum for the methanogens. The gas content of the sediments would thus increase with depth as reducing conditions became more favorable but would then become depleted as higher temperatures hindered microbial growth. Shipboard chromatographic analysis of evolved gases revealed a preponderence of methane which was also consistent with a biogenic source.

Subsequent laboratory analyses of the gases from Leg 10 by Claypool et al. (1973) suggest that cores from Site 88 may contain some thermogenic gas. Results of the gas chromatography (GC) and ^{13}C isotope analysis are summarized in Table 3. Differences can be noted between values obtained from Site 88 and those from Sites 90 and 91. The deeper cores from Site 88 show a much greater abundance of hydrocarbons heavier than C_2, while those from Sites 90 and 91 are almost pure methane. The ^{13}C isotopic content of methane from Site 88 has a mean $\delta^{13}C$ of -54.7. This is much lighter than the mean $\delta^{13}C$ of 80.6 for Sites 90 and 91. Both isotopically light methane and the presence of significant amounts of C_2 - C_4 hydrocarbons indicate gas produced during thermal maturation of kerogen (Rice and Claypool, 1981). These two parameters are cross plotted in Figure 39. The samples from Site 88 form a separate group, distinct from the clearly biogenic samples from Sites 90 and 91.

Sources of Thermogenic Gas

Using seismic stratigraphy and thermal modeling it is possible to ascertain the probable depths of thermally mature source rocks capable of having generated the gas contained in the thermogenic hydrates recovered from the Green Canyon area and those inferred to exist at DSDP Site 88.

Rich, mature source rocks are abundant over wide areas of the Gulf of Mexico region. The prodigious amounts of petroleum and gas that have been produced throughout onshore and shallow marine areas suggest widespread mature source beds with accumulation of petroleum being controlled by migrational and trapping parameters. The rocks which were sampled by drill holes in water deep enough to stabilize gas hydrates were not buried deeply enough or for sufficient time to have attained thermal maturity. Thus, the seaward extent of the rich source beds underlying the continental shelf petroleum province must be deduced by more indirect means.

In a study of oil and gas potential in the deep portions of the study region, Foote et al. (1983) speculated that Upper Jurassic organic-rich shales equivalent to those found on the Mexican coastal plain may be widespread enough to be a major source of oil. These units may be correlative with the Upper Jurassic Smackover formation which has high source potential to the north of the study region (Tolson et al., 1982, in Foote et al., 1983). Intervening salt piercement structures preclude long-range correlation of the Smackover into deeper water.

Seismic stratigraphy of the Sigsbee Abyssal Plain suggests the presence of more recent potential source beds throughout the western Gulf region. Although areas of interest for gas hydrates include the continental slope and rise areas in addition to the abyssal plain, stratigraphy of the slope and rise is severely disrupted by widespread salt diapirism. The stratigraphic configuration of the slope and rise areas which controlled the maturation history of potential source beds is best appraised by examining the undisturbed strata of the abyssal plain and reconstructing the possible slope and rise stratigraphy by extrapolation. Seismic stratigraphic studies of the deep basin

TABLE 3

CHEMICAL AND ISOTOPIC COMPOSITION OF DSDP GASES FROM THE GULF OF MEXICO. AFTER CLAYPOOL ET AL., 1973.

Sample	Depth Below Sea Floor (m)	Component (Volume %)			δC^{13}(°/..) CH_4
		CH_4	C_2H_6	CO_2	
88-2	54	99.5	0.06	0.42	-59.0
3	102	96.6	3.03	0.32	-54.7
4-4	110	91.5	8.39	0.09	-48.7
4-6	113	97.0	2.67	0.30	-54.8
5	135	95.9	3.97	0.17	-56.3
90-3	130	99.4	n.d.	0.63	-83.0
11	678	99.9	0.01	0.14	-78.0
91-3	159	99.8	n.d.	0.19	-84.0
5	190	99.6	n.d.	0.45	-80.4
6	301	99.1	0.01	0.89	-81.0
7	412	99.7	0.01	0.32	-80.8
8	492	99.1	0.02	0.85	-80.0
11	774	99.3	0.01	0.72	-79.9
18	838	99.2	0.02	0.82	-78.7

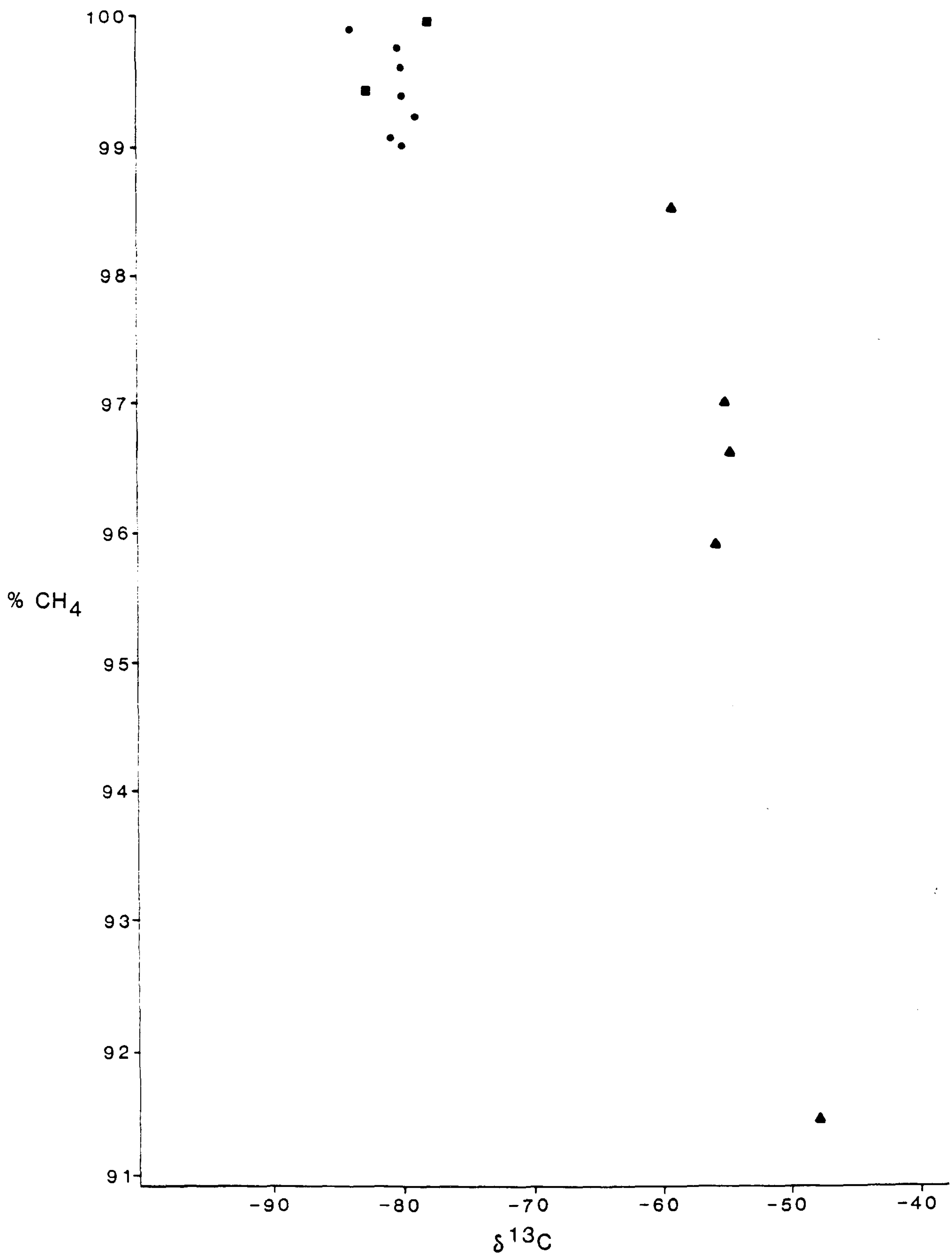


Figure 39. METHANE CONTENT AND ISOTOPIC COMPOSITION OF GASES FROM DSDP SITES 88 (▲), 90 (■), AND 91 (•)

of the Gulf, the Sigsbee Abyssal Plain, have followed the terminology introduced by Watkins et al. (1976; Table 2). A more recent treatment by Shaub et al. (1984) indicates that potential source lithologies may be widespread throughout the basin. The basal sedimentary unit, the Challenger, has been interpreted as consisting of 2,000 to 3,000 m of evaporite beds and marine clastics of Late Jurassic to middle Cretaceous age. Evaporites have been demonstrated to have source potential in other regions (Kirkland and Evans, 1981). Some units within the clastic sequences probably contain sufficient organic matter for thermogenic gas generation.

The overlying Campeche unit consists of 1,000 to 3,000 m of middle Cretaceous to possibly Eocene deep marine rocks. It is reasonable that the Campeche unit would contain organic-rich beds, especially if the Cretaceous anoxic events of epeiric seas also affected deeper basins such as this. The Lower and Upper Mexican Ridges seismic units record Eocene through Miocene deposition. The inferred lithologies of these units -- fine-grained, deep marine clastics -- could have source potential, although the thermal history may not have been adequate for maturation. The Miocene through Holocene Cinco de Mayo and Sigsbee units are too shallow to be considered as catagenetic source rocks.

A rough estimate of which stratigraphic intervals have been sufficiently heated for hydrocarbon generation can be obtained by burial history reconstruction. Although many modeling techniques are in use, Lopatin's method as described by Waples (1980) is used in this report. Briefly, the method assumes that organic matter matures linearly with time and at a rate which doubles for each 10° change in temperature. A value to measure maturity, termed the time - temperature index (TTI) is derived. The "oil generation window" of maturity corresponds to TTI values of 15 to 160. The upper limit of preservation of gas composed of methane and significant C_2 - C_4 hydrocarbons is about TTI = 1,500.

A simple reconstruction was calculated using thermal data and sedimentation rates from Epp et al. (1970). Present-day geothermal gradients in the deep reaches of the Gulf average 4.0°C/100 m. A mean bottom water temperature of 4.0°C was measured. Epp et al. derived a mean sedimentation rate of 5 to 10 cm/1,000 years for the Campeche Gulf. TTI values were calculated for 5 and 10 cm/1,000 years and assuming a constant geothermal gradient through time. The results are presented in Figures 40 and 41. Since the thermogenic gas hydrates thus far discovered have been associated with petroleum stained sediments, it is reasonable to assume that the gas was generated at depths at least as great as was the oil. Thus, the oil generative window (TTI = 15 to 160) is plotted on Figures 40 and 41. Gas of similar composition to that found in thermogenic gas hydrate cores can also result from decompo- sition of kerogen and bitumen at temperatures higher than those necessary for oil generation. Therefore the lower limit of "wet" gas occurrence (TTI = 1,500) is also plotted on the diagrams. The reconstructions indicate that Upper Cretaceous through middle Tertiary rocks would have the proper maturity for thermogenic gas generation. Which age rock is the most likely source is strongly dependent on the sediment accumulation rate selected for the model, illustrating a weakness of such a general approach. Regardless of the assumed sedimentation rate, the models are in agreement in that thermogenic gas must have migrated at least 2,000 m vertically from the shallowest mature source beds to the base of the gas hydrate formation zone.

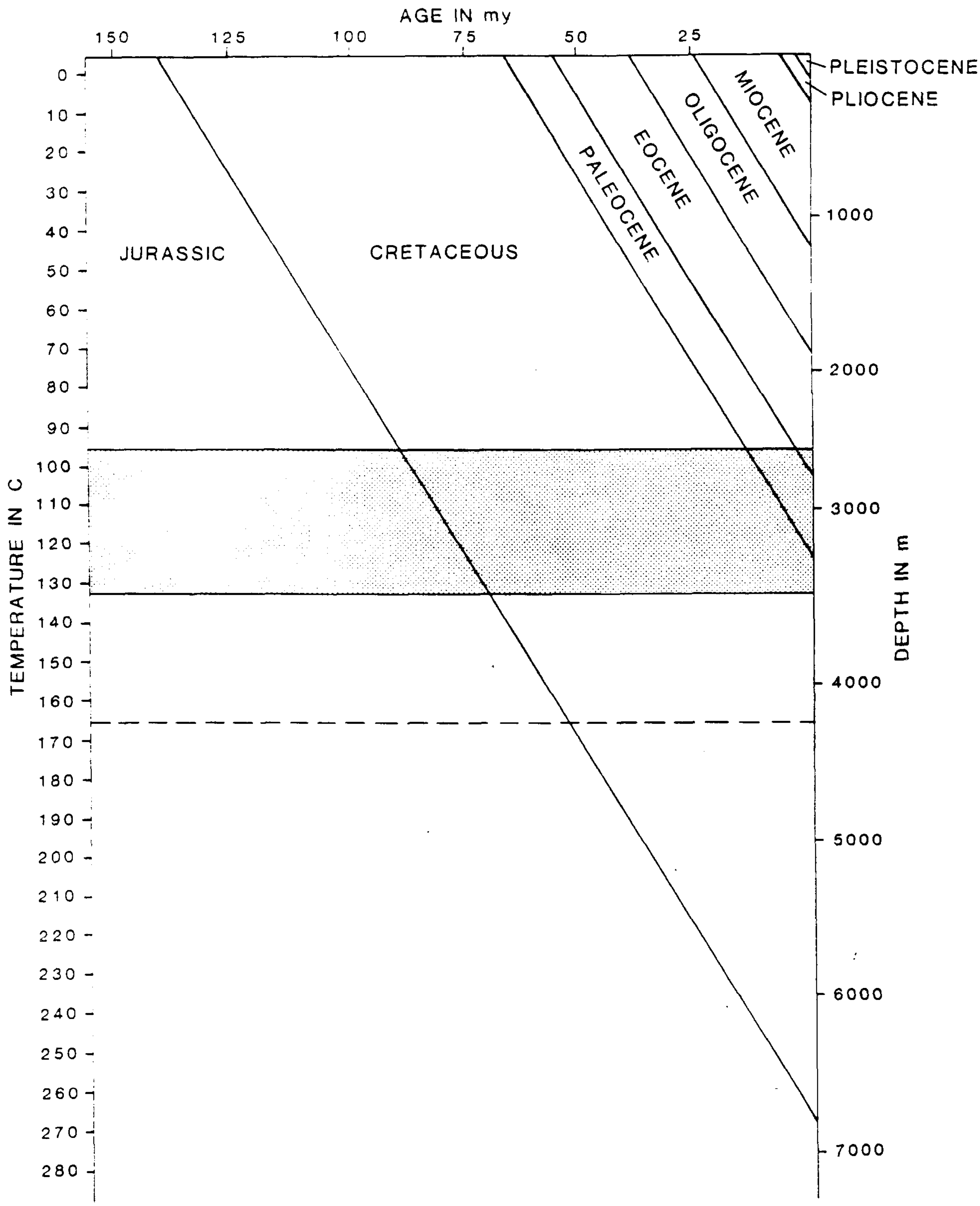


Figure 40. **LOPATIN BURIAL HISTORY RECONSTRUCTION OF DEEP GULF OF MEXICO, ASSUMING SEDIMENT ACCUMULATION RATE OF 5cm/1000yr**

Geothermal gradient = 4°C/100m. Patterned area represents main zone of thermogenic hydrocarbon production (TTI = 15 to 160); dashed line indicates deepest occurrence of "wet" gas (TTI = 1500) according to Waples (1980).

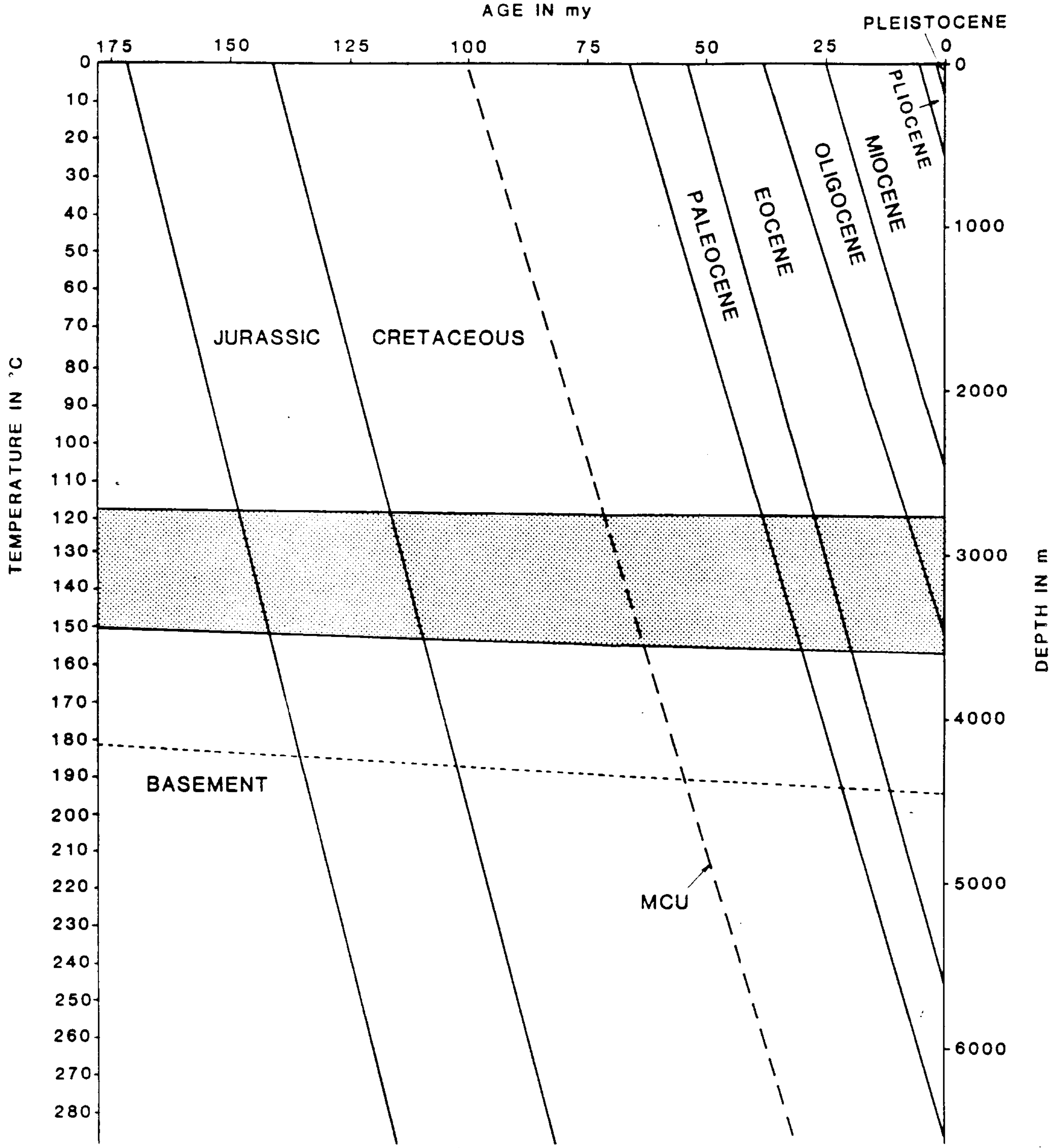


Figure 41. **LOPATIN BURIAL HISTORY RECONSTRUCTION OF DEEP GULF OF MEXICO, ASSUMING SEDIMENT ACCUMULATION RATE OF 10cm/1000yr**

Geothermal gradient = 4°C/100m. Patterned area represents main zone of thermogenic hydrocarbon production (TTI = 15 to 160); dashed line indicates deepest occurrence of "wet" gas (TTI = 1500) according to Waples (1980). MCU - Middle Cretaceous unconformity.

Such a long migration distance through fine-grained rocks indicates that structural migration pathways such as faults and associated fracture zones are necessary.

DSDP drilling and seismic stratigraphy permit refinement of the previous burial history diagram. Precise dating of DSDP cores by foraminifera and palynomorphs permitted more accurate estimates of the sedimentation rates of drilled intervals. Improved coverage of deep areas of the Gulf region by seismic lines and reliable estimation of the seismic velocities of deep sedimentary units have yielded isopach maps of sedimentary intervals (Shaub, 1984; Figures 8 - 13, Table 1). From these maps sedimentation rates can be estimated for a particular site.

A burial history reconstruction for DSDP Site 91 on the Sigsbee Abyssal Plain is summarized in Figure 42. The drill reached a subbottom depth of 890 m, beneath the middle - upper Miocene boundary. Since the rock units which constitute the seismic stratigraphic units older than middle Miocene have not been drilled, precise ages are not known. Estimates of boundaries used for thermal modeling are:

Upper Mexican Ridges - Cinco de Mayo: 10 m.y.

Lower Mexican Ridges - Upper Mexican Ridges: 33 m.y.

Campeche - Lower Mexican Ridges: 55 m.y.

Challenger - Campeche: 100 m.y. (Mid-Cretaceous Unconformity)

Crust - Challenger: 170 m.y.

The reconstruction (Figure 42) bears a strong resemblance to Figure 40 which was made without stratigraphic data and based only on a rough sedimentation rate estimate of 5 cm/1,000 yr. That estimate was close to the 5.2 cm/1,000 yr. rate of accumulation above the middle Cretaceous unconformity which separates the Challenger and Campeche Units.

If the boundary hypotheses are reasonable, other inferences can be made from this more detailed model. The main zone of oil formation is found at a present depth of 2,800 to 3,800 m subbottom suggesting that recently migrated thermogenic hydrocarbons would most likely have come from the Upper Cretaceous through Paleocene Campeche unit with possible minor contributions from the lower Eocene portion of the Lower Mexican Ridges unit. The deepest calculated occurrence of wet thermogenic gas at this site is around 4,500 m subbottom. That implies a vertical migration distance of 2,000 to 4,000 m from potential source beds to the gas hydrate stability zone.

Which units the hydrocarbons would be derived from could vary depending on time of primary migration. Foote et al. (1983) cite evidence of Pleistocene diapiric activity. Worzel and Burke (1979) describe seismic sections from intraslope basins that strongly suggest salt movement in the early Tertiary. If fracture systems resulting from this structural deformation are the migration pathways for the thermogenic gas, early Tertiary primary migrations of hydrocarbons should have involved Jurassic to Lower Cretaceous rocks of the Challenger unit.

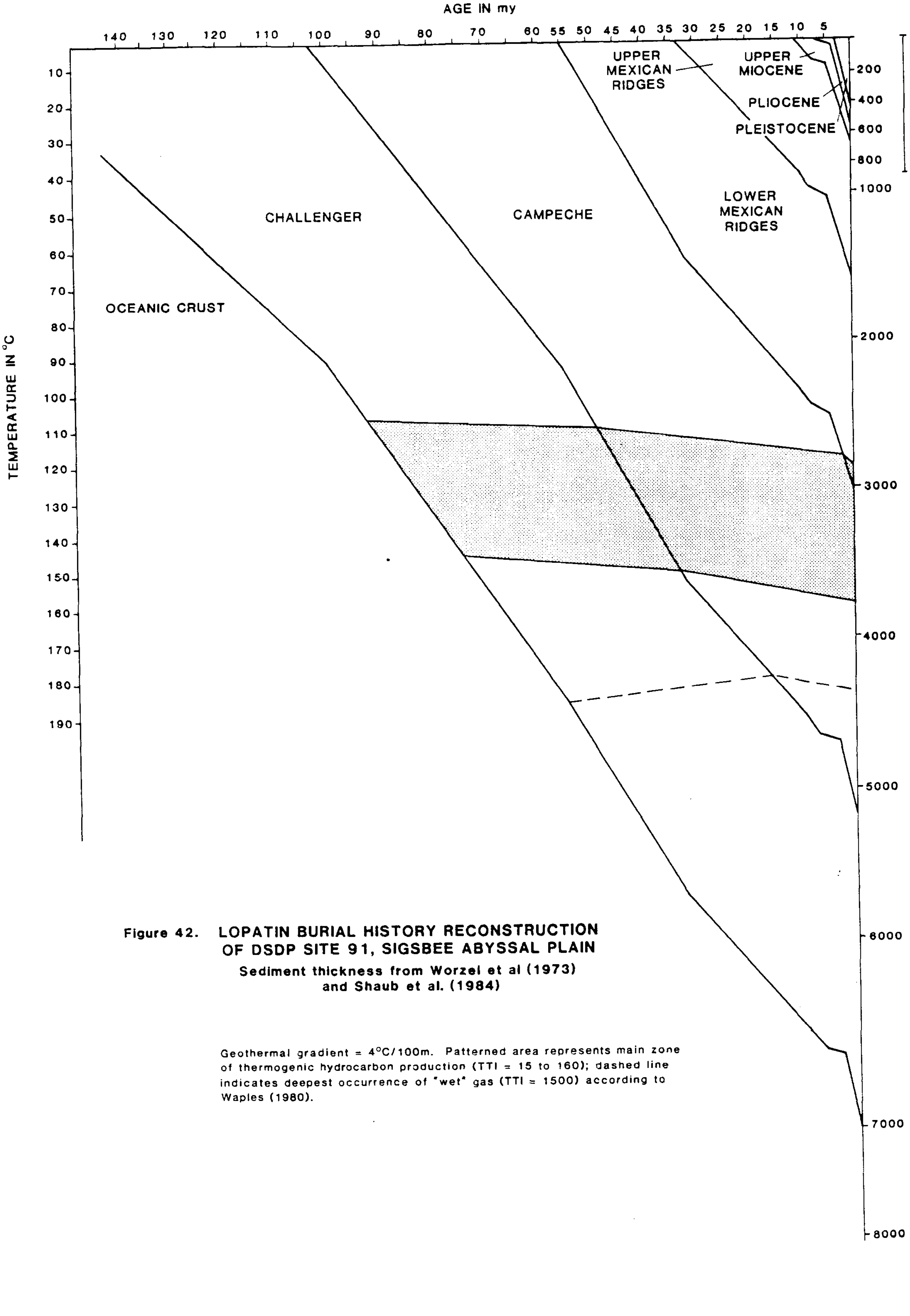


Figure 42. LOPATIN BURIAL HISTORY RECONSTRUCTION OF DSDP SITE 91, SIGSBEE ABYSSAL PLAIN
Sediment thickness from Worzel et al (1973) and Shaub et al. (1984)

Geothermal gradient = 4°C/100m. Patterned area represents main zone of thermogenic hydrocarbon production (TTI = 15 to 160); dashed line indicates deepest occurrence of "wet" gas (TTI = 1500) according to Waples (1980).

Similar Lopatin plots were constructed to extend the above findings on thermal maturity from the Sigsbee Abyssal Plain to areas deformed by salt diapirism (Figures 43 and 44). Figure 43 graphs the thermal history of sediments near the Challenger Knoll in the Sigsbee Salt Dome province. The thickness values are derived from maps by Shaub et al. (1984; Figures 8 - 13) for a site slightly offset from Challenger Knoll itself. Hydrocarbons were recovered from Challenger Knoll by coring at DSDP Leg 1, Site 2, indicating active migration. The Lopatin diagram for Challenger Knoll (Figure 43) strongly resembles that of the abyssal plain, and the above conclusions of probable source beds and migration distances should apply to both areas.

The implication of the presence of thermogenic gas hydrates on a salt structure in the Campeche Knolls (DSDP Site 88) dictated an examination of the time - temperature relationships at that site. The burial history reconstruction (Figure 44) shows little divergence from the two previous plots.

In summary, TTI plots of the deep gulf indicate large-scale vertical migration of gas is necessary to form thermogenic hydrates. The probable age of the actual source beds varies with the time of primary migration. With early Tertiary migration gas should have come from the Jurassic to Lower Cretaceous Challenger unit. Pleistocene migration requires younger source rocks, probably Upper Cretaceous through Eocene Campeche or Lower Mexican Ridges units. The assumptions of constant geothermal gradient through time should not grossly distort these results, especially if the thermal perturbations associated with diapirism were limited to the Quaternary.

The source of oils produced from the continental slope areas of offshore Louisiana and associated thermogenic gas hydrates is difficult to ascertain because of large-scale disturbance of source beds by diapirs. Some work was done in the less complicated area closer to shore which may be extrapolated into deeper slope areas where hydrates may be stable. Dow (1977) reconstructed the burial history of offshore Louisiana in less than 200 m of water from results of 12 wells. Since very rapid sedimentation occurred in the Tertiary and Quaternary due to outbuilding of the Mississippi Delta, the depths to mature source rocks would be expected to be greater and the ages of the source rocks would be expected to be younger than in the more slowly filled deep slope and abyssal basin areas.

Dow's reconstruction indicates that 5,000 to 7,000 m of vertical migration would be necessary for gas to reach the sea floor. However, if extrapolated to the slope area with less rapid sedimentation, 3,000 m migration seems a reasonable estimate. Geochemical work by Young et al. (1977) indicated a typical migration distance of 3,300 m for light hydrocarbons in the offshore Louisiana continental shelf and upper slope.

Regardless of the model employed to calculate thermal maturity, the use of reasonable values for sedimentation rates and geothermal gradients indicates that vertical migration of 2,000 to 3,000 m is required for thermogenic hydrates. This suggests that the most probable locations for thermogenic gas hydrates are adjacent to or over structural features which may provide permeable conduits for gas migration of the scale required.

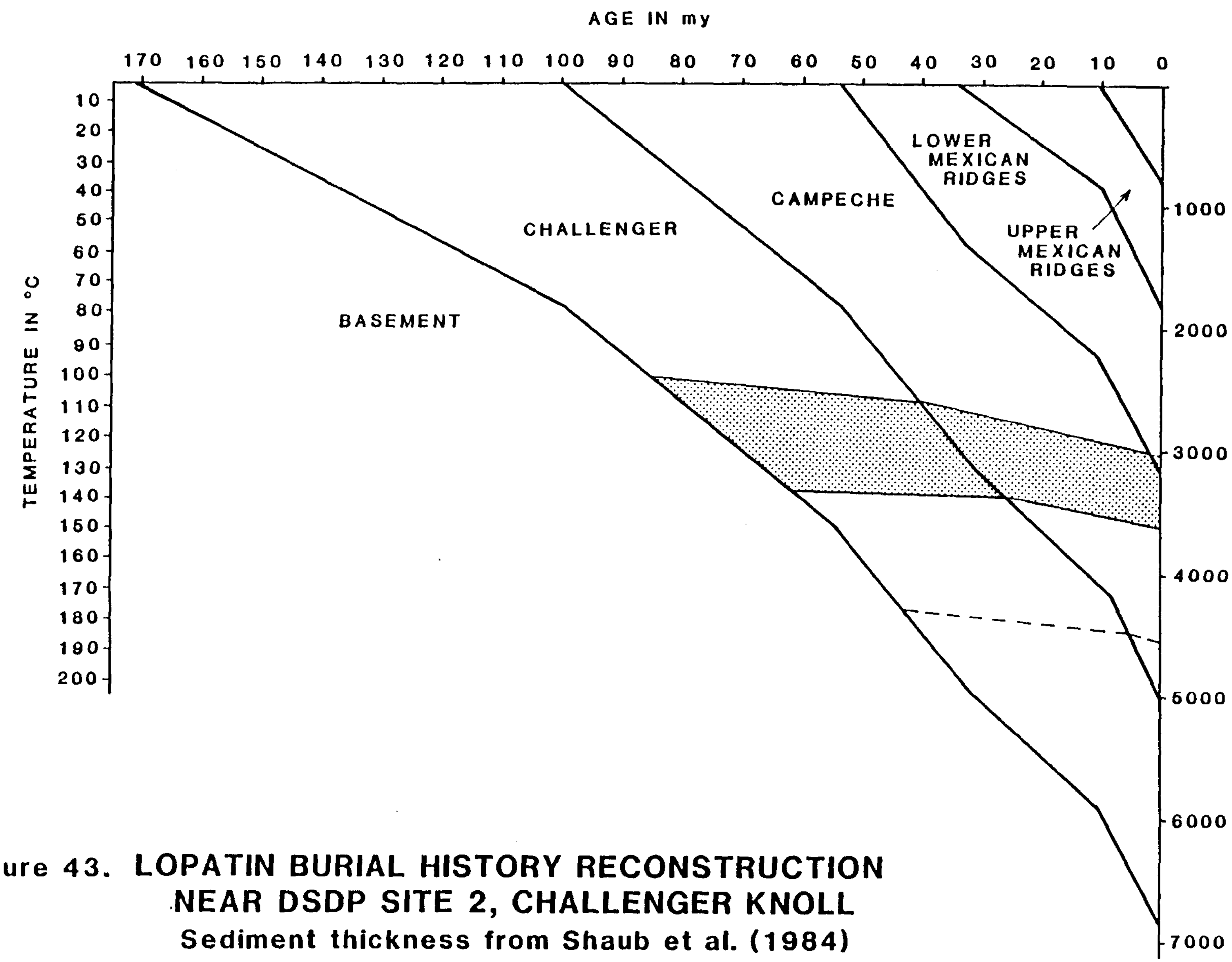


Figure 43. **LOPATIN BURIAL HISTORY RECONSTRUCTION NEAR DSDP SITE 2, CHALLENGER KNOLL**
Sediment thickness from Shaub et al. (1984)

Geothermal gradient = 4°C/100m. Patterned area represents main zone of thermogenic hydrocarbon production (TTI = 15 to 160); dashed line indicates deepest occurrence of "wet" gas (TTI = 1500) according to Waples (1980).

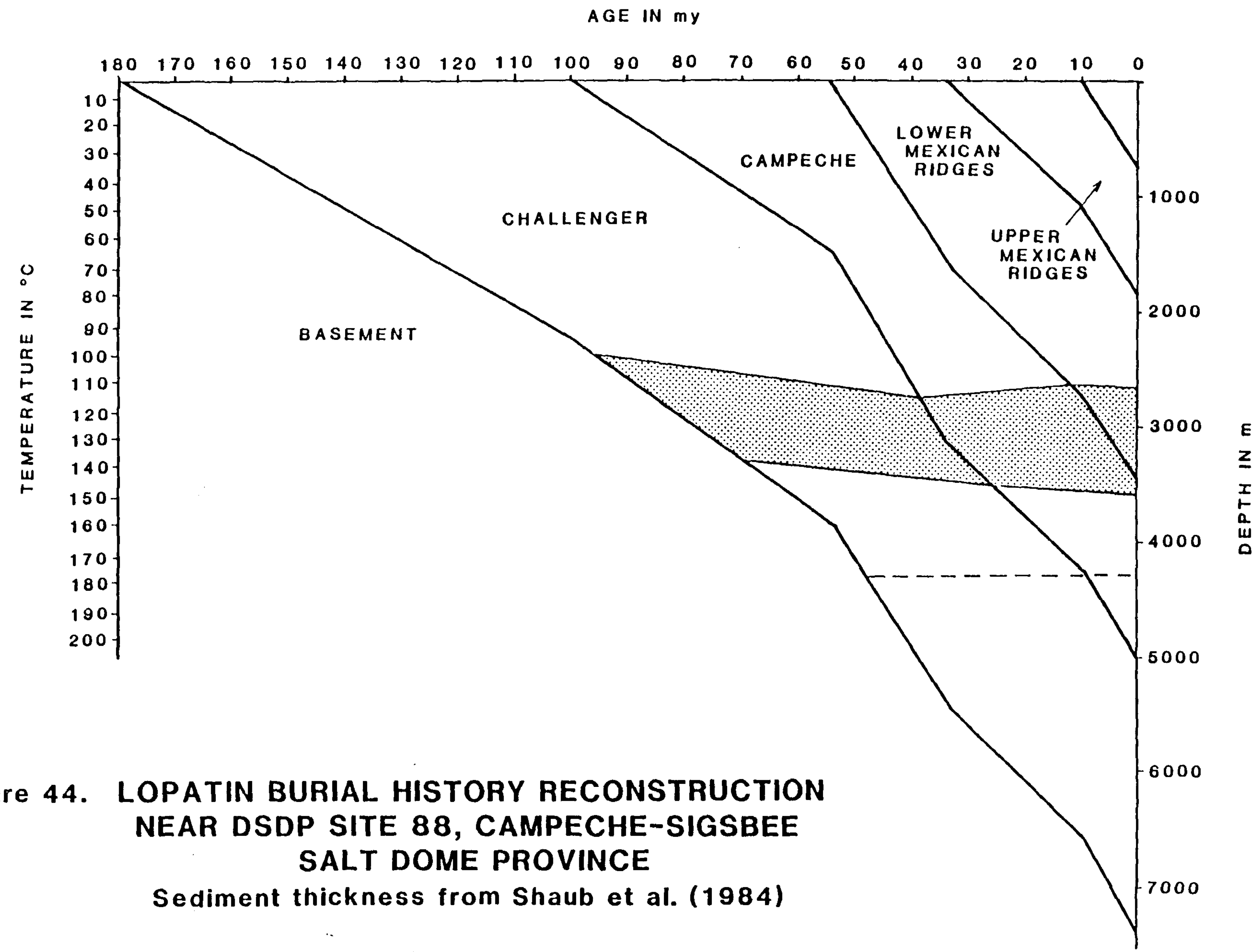


Figure 44. **LOPATIN BURIAL HISTORY RECONSTRUCTION NEAR DSDP SITE 88, CAMPECHE-SIGSBEE SALT DOME PROVINCE**

Sediment thickness from Shaub et al. (1984)

Geothermal gradient = 4°C/100m. Patterned area represents main zone of thermogenic hydrocarbon production (TTI = 15 to 160); dashed line indicates deepest occurrence of "wet" gas (TTI = 1500) according to Waples (1980).

Structural Control of Thermogenic Gas Hydrates

The migration distances necessary for thermogenic gas hydrates indicate that substantial accumulations of thermogenic gas hydrates should be limited to structurally disturbed areas. As previously described in this report, the principal structures which could disrupt the sedimentary column to produce a migrational pathway in the study regions are growth faults, thrust faults associated with folding on the Mexican Ridges, and salt and shale diapirs.

To date, all proven or inferred thermogenic gas hydrate locations have been near salt diapirs. Brooks and Bryant (1985b) have mapped the tops of the salt structures near the two sites where thermogenic gas hydrates have been recovered. Recovery sites in lease blocks 184 and 190 are plotted on their map in Figure 45.

Using a representative seismic velocity for water (1,480 m/sec) and shallow sediments (2,400 m/sec; Shaub et al., 1984) the time contours of Figure 45 can be used to estimate depths to salt structures. Site 184 is located about 5 km north, 5 km southeast and 8 km southwest of large salt diapirs. The ocean depth at 184 is 560 m (Brooks and Bryant, 1985a) or .71 seconds. The salt is about 2,500 m below the ocean floor ((2.8 sec - 0.7 sec) x 1,200 m/sec). The adjacent salt structures are at most 360 m subbottom depth ((1 sec - 0.7 sec) x 1,200 m/sec). Site 190 is located in 560 m (0.8 sec) of water. It is located about 500 m above the salt ((1.5 - 0.8) sec x 1,200 m/sec). Three diapirs with depths of less than 240 m surround the site at distances of 5, 4, and <1 km. These gas hydrate bearing sites are each on the flanks of diapirs or in intraslope basins. Lack of more extensive data precludes comparison of the structural settings of thermogenic hydrates with those of nearby biogenic gas hydrate sites in the northwestern Gulf slope area.

The only inferred occurrence of thermogenic gas hydrates in the deeper southern area of the Gulf study region was also associated with a salt diapir. DSDP Site 88 was drilled directly on a salt diapir (Figure 46), with a relief of over 800 m above the sea floor, with probably 350 m of sediments over the salt core (Worzel et al., 1973).

Comparison of Site 88 with other DSDP holes drilled over salt structures shows that presence of a diapir does not guarantee a thermogenic hydrate accumulation in the Gulf. DSDP Site 2 on the Challenger Knoll in the Sigsbee Plain (Figure 47) recovered Miocene through Pleistocene sediments which contained residual petroleum. The hole bottomed in Jurassic dolomite - anhydrite cap rock of an underlying salt dome. Abundant gas which was produced while drilling caused termination of the drilling before the salt target was reached due to concerns over a possible escape of gas into the environment. Although there exists no evidence that thermogenic gas hydrates were found at Site 2, it is possible that some hydrates were drilled through. Hydrates were inferred at nearby Sites 88 - 91 based on unusual degassing characteristics. At Site 2, conditions indicative of gas hydrate dissociation may not have been noticed; being only the second hole drilled, no standard of comparison for deep sea sediments existed to determine what constitutes abnormal degassing. Indeed, given the depth and temperature conditions, and the abundant thermogenic gas encountered in drilling through the hydrate stability zone, the existence of gas hydrates on Challenger Knoll seems plausible. Another salt structure was drilled on Leg 10 at Site 92 on the Sigsbee Scarp about 300 km north of Site 2. The hole was drilled through

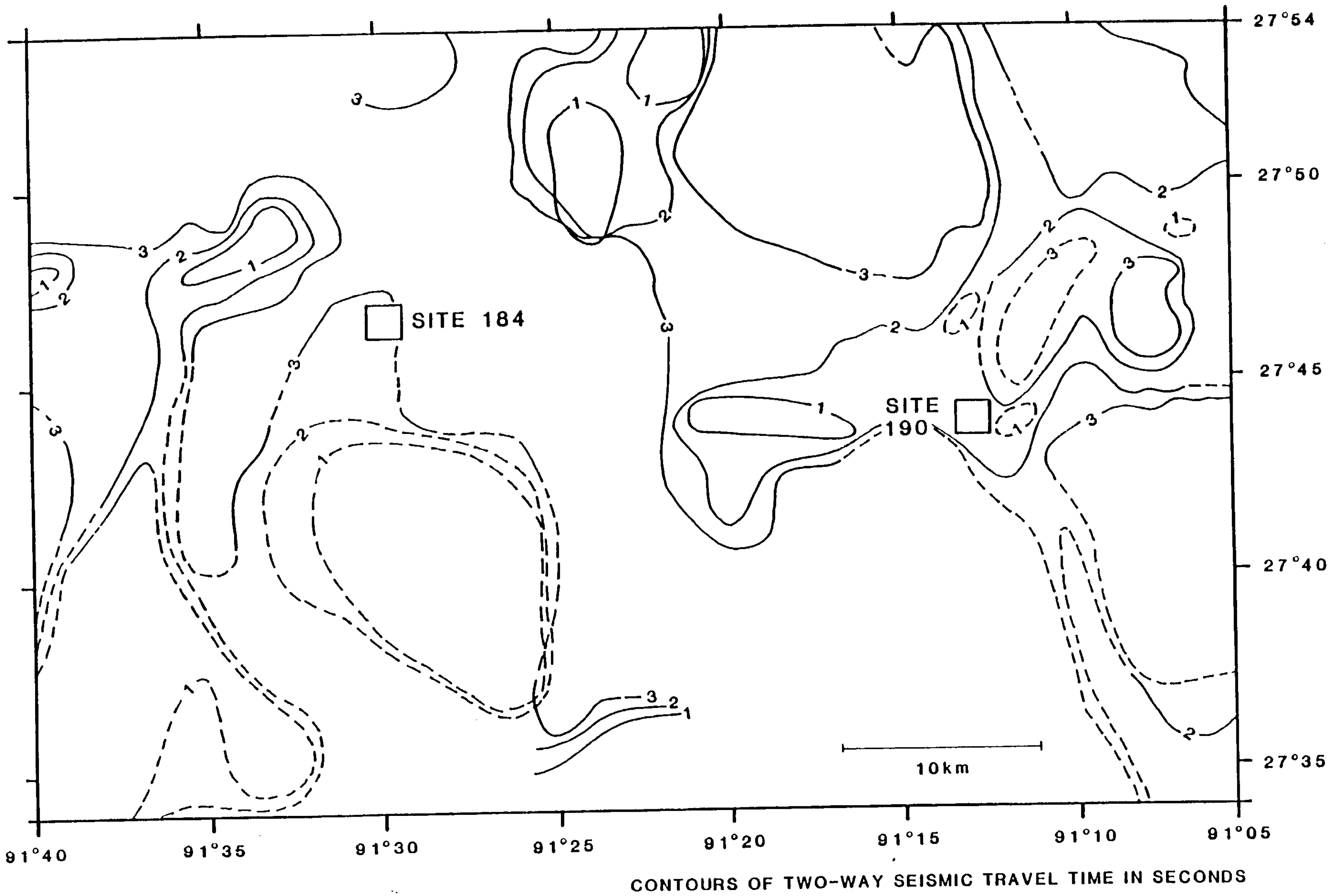


Figure 45. **STRUCTURE CONTOUR MAP OF TOP OF SALT IN AREA OF THERMOGENIC GAS HYDRATE RECOVERY**

After Brooks (1985b)

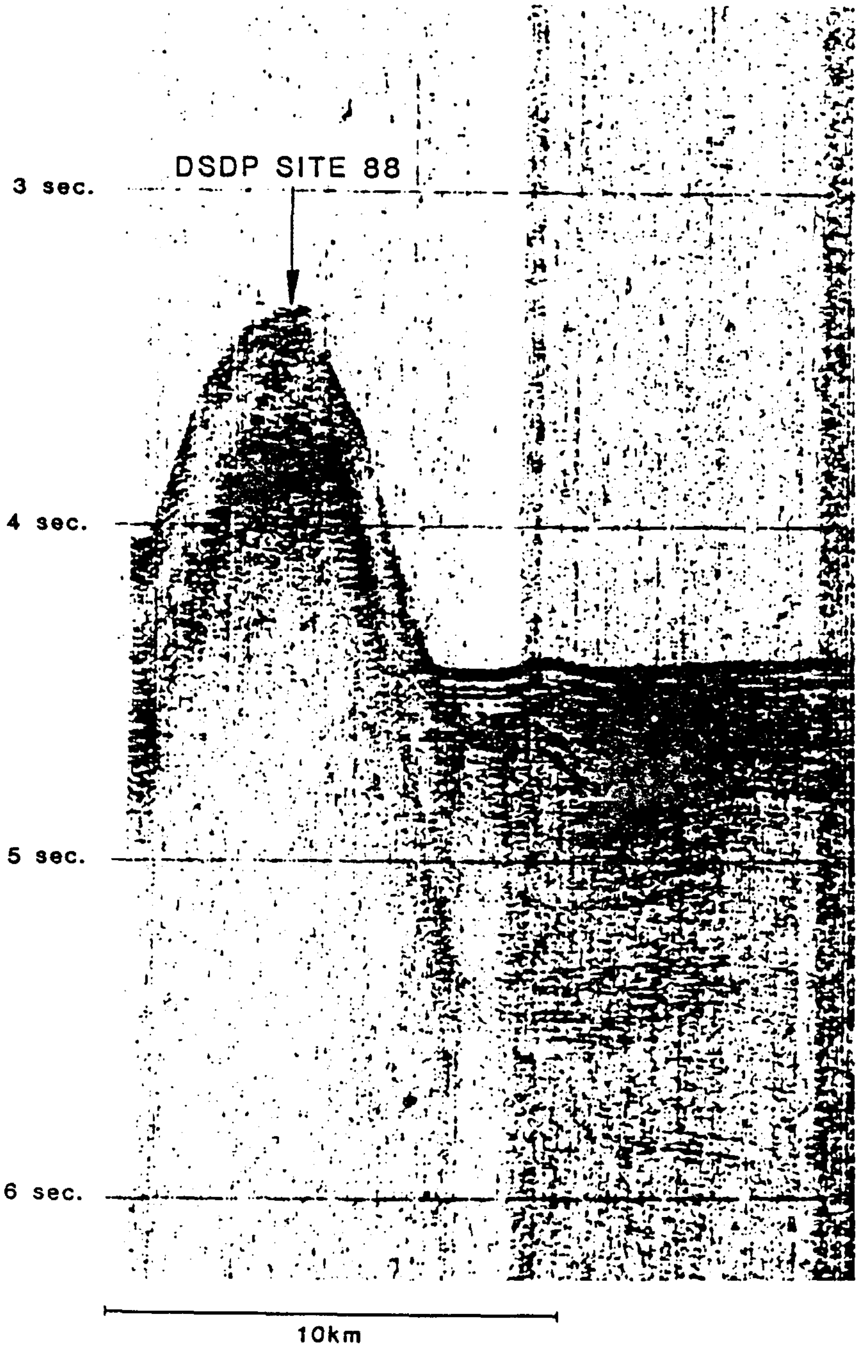


Figure 46. SEISMIC PROFILE OF DSDP SITE 88, CAMPECHE KNOLLS

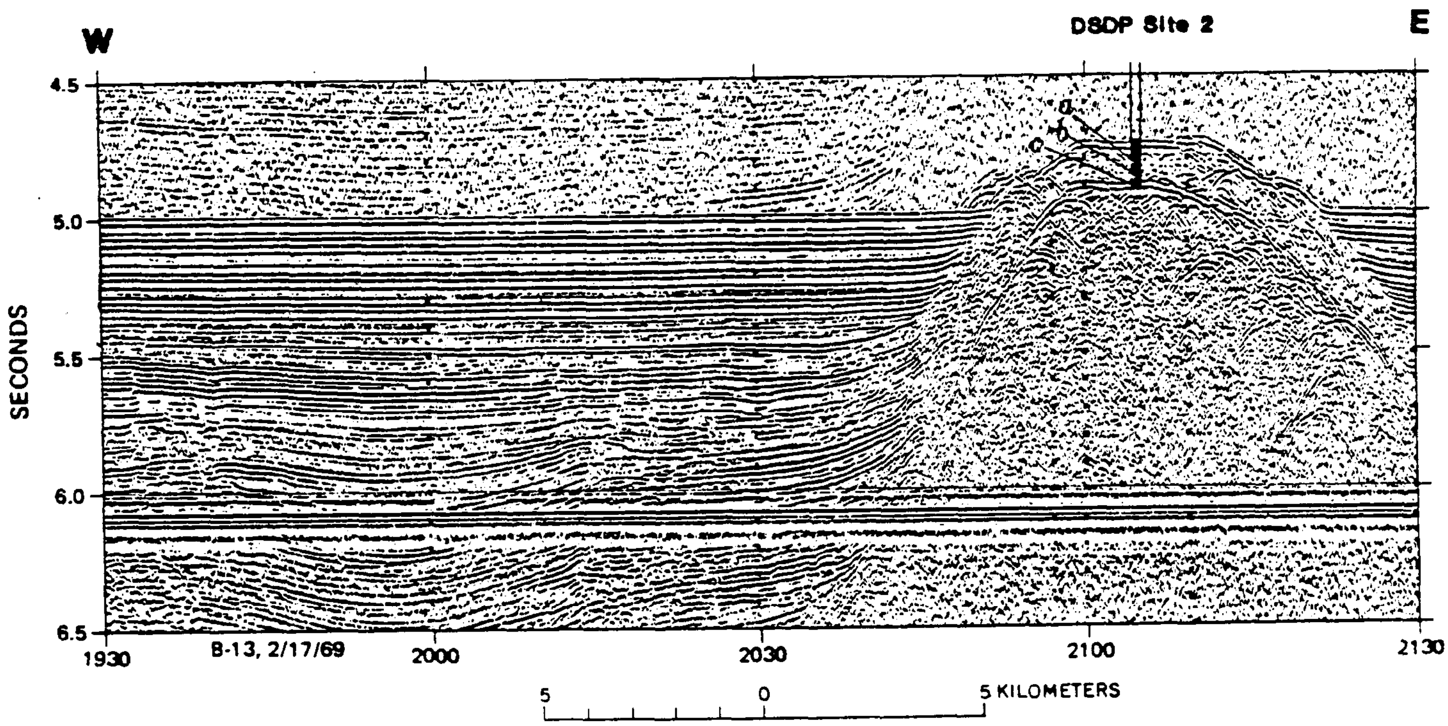


Figure 47. SEISMIC PROFILE OF CHALLENGER KNOLL

After Garrison and Martin (1973).

282 m of Pleistocene ooze and fine-grained turbidites without encountering significant gas. No unusual core degassing features were noted.

Thus, structural disturbance due to diapirism may be a necessary but not sufficient condition for thermogenic gas hydrate formation. Traps are formed concurrently with migrational pathways during diapiric uplift. Thus, highly efficient hydrocarbon traps along the flanks may have prevented sufficient gas leakage to the hydrate zone for significant gas hydrate formation. The thermogenic gas hydrates in the northern Gulf margin area occur in close proximity to active gas and oil seeps (Brooks and Bryant, 1985a). This suggests that structurally disturbed areas devoid of gas hydrates may also represent sites where efficient upward migration of gas has taken place when shallower water or warmer temperatures may have prevented hydrate formation. Such migration may have exhausted source hydrocarbons to the point that insufficient upward flux of gas exists in recent times to supersaturate the pore water to the point that hydrates nucleate.

Diapirism may affect gas hydrate formation in ways other than by providing migration routes. The high thermal conductivity of salt, along with its physical displacement sediments results in increased heating of shallow sediments. Keen (1983) has shown that diapirism can increase thermal maturity of shallow sediments, principally by introduction of a high temperature hydrothermal system adjacent to the diapirs. Thus, shallow sediments near diapirs may be sufficiently mature to generate hydrocarbons which would allow shorter migration distances for gas incorporated into hydrates. A minor factor that may come into play is dragging of deep, mature source rocks to shallower depths along the flanks of the diapir which may similarly reduce the necessary migrational distances for thermogenic gas hydrate formation.

Stability of Thermogenic Gas Hydrates

Unique properties of thermogenic gas hydrates and the sedimentary environments in which they are found may affect their stability relative to biogenic gas hydrates.

Saline pore waters near salt diapirs may destabilize thermogenic gas hydrates. Pore water at DSDP holes drilled over salt structures (Sites 2, 88, and 92) showed increased salinity with depth (Figure 48). This was attributed to increasing proximity to underlying salt with depth. Salt is an effective inhibitor of gas hydrates, occasionally being used industrially for that purpose (Makogon, 1978). Data from a chart by Scott et al. (1980) on the inhibition of hydrate formation by sodium chloride (Figure 49) shows that salt is a greater destabilizing influence on thermogenic gas hydrates (0.68 gravity gas) than on biogenic (methane) gas hydrates at typical seawater salinity (3.6%). The formation temperature of methane hydrates is depressed 0.4°C by typical seawater salinity relative to pure water. The temperature of gas hydrate stability for thermogenic gas (0.68 gravity) is depressed 1.2°C by typical seawater compared to pure water. The depression of gas hydrate formation temperature and the resultant thinning of the gas hydrate zone due to increased interstitial salinity is presented in Table 4 for DSDP Sites 2, 88 and 92 based on chemical data from Kaplan and Presley (1969) and Manheim et al.

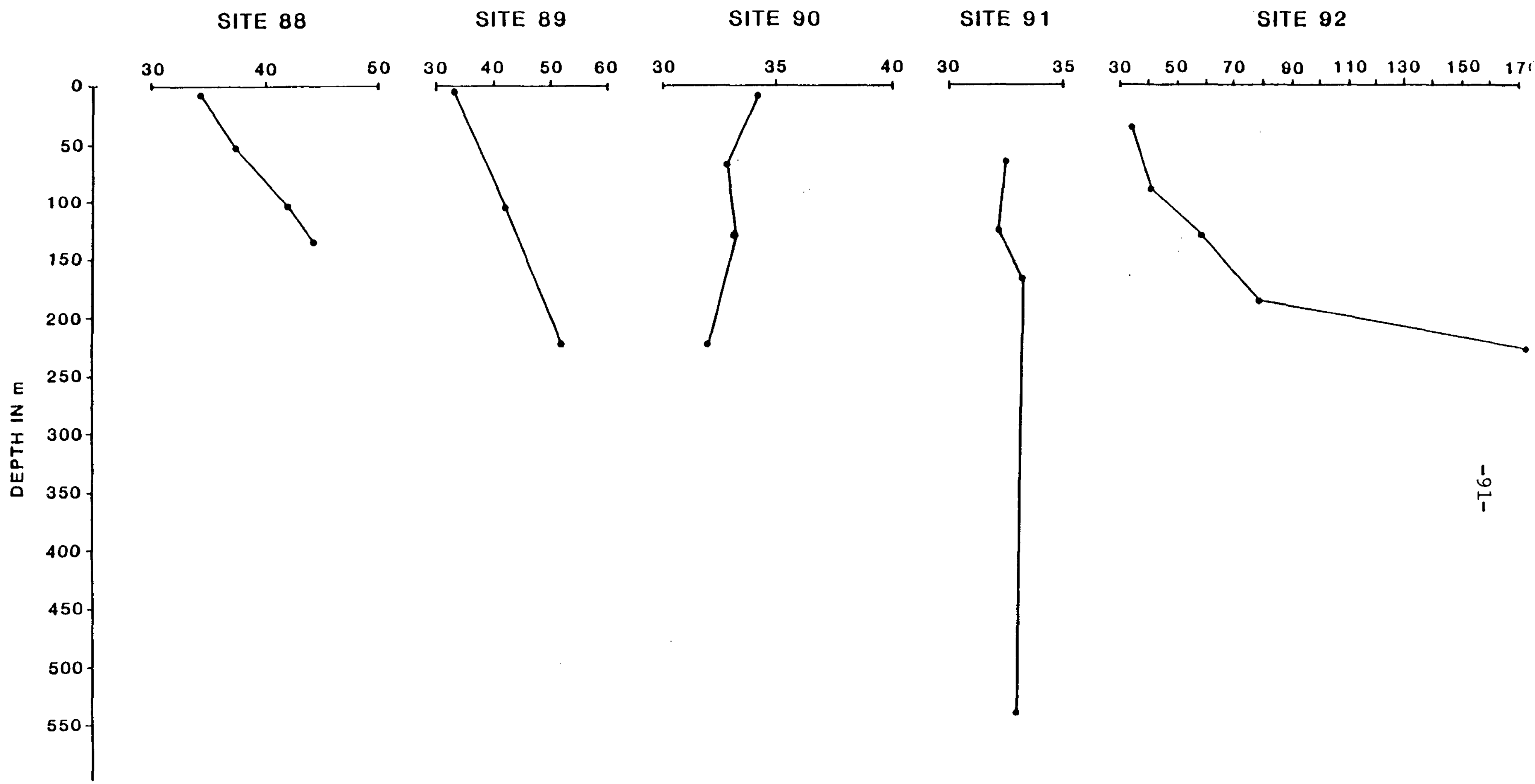


Figure 48. **PORE WATER SALINITY OF SELECTED DSDP SITES IN GULF OF MEXICO**

Data modified from Manheim et al. (1973) to NaCl Equivalent by method of Makogon (1978). Salinity in parts per thousand

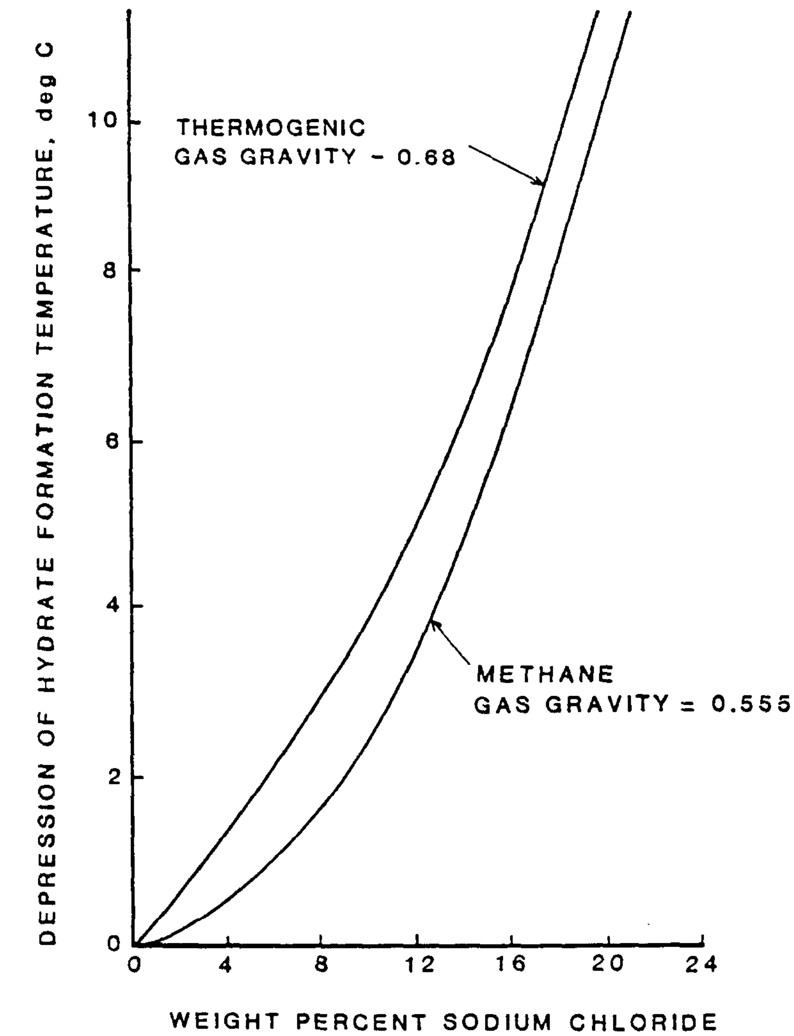


Figure 49. DEPRESSION OF GAS HYDRATE FORMATION TEMPERATURE BY SODIUM CHLORIDE

After Scott (1980)

TABLE 4.

EFFECT OF SALINE PORE WATER ON GAS HYDRATE STABILITY

DSDP Site	Max Salinity (corrected) (%)	Biogenic Hydrate: Change in Equil. Temp.(°C)	Biogenic Hydrate: Change in Hydrate Depth (m)	Thermogenic Hydrate: Change in Equil. Temp.(°C)	Thermogenic Hydrate: Change in Hydrate Depth (m)
2	6.5	-0.74	18.5	-1.60	40
88	4.4	-0.30	7.5	- .45	11
89	5.2	-0.40	10.0	- .70	18
92	17.0	-6.46	161.0	-7.22	181

(1973). The reported values for Na^+, K^+, Mg^{2+}, Ca^{2+}, and Cl^- were corrected to obtain a NaCl concentration equivalent in gas hydrate inhibition activity to that of the solution of mixed ions using the method of Makogon (1978). The NaCl equivalents for maximum pore water salinity for each hole thus obtained were then plotted on Figure 49 to obtain the values in Table 4. Significant destabilization of gas hydrates is seen for Site 92, where no hydrates were found. If the increased salinity also affects the degree of supercooling necessary to initiate hydrate growth, then net effect may be more pronounced than that indicated in Table 4 for gas hydrate equilibrium temperatures.

Increased heat flow near a diapir may diminish the gas hydrate stability zone in nearby sediments in a straightforward manner.

The presence of ethane, propane, and iso-butane stabilizes thermogenic gas hydrates. Using a pressure - temperature relationship developed from the data of Holder (1983) and John (1981) to determine thickness and depth of gas hydrates, thermogenic gas hydrates would have a stability zone about 200 m thicker and deeper in the sediment than biogenic gas hydrates under similar conditions. Thermogenic gas hydrates could therefore be expected to be stable in shallower water than biogenic hydrates.

The presence of oil in sediments containing thermogenic gas hydrates complicates stability relationships. Oil such as that reported staining cores of gas hydrate bearing sediments introduces another component and another phase into the thermodynamics of hydrate formation. The solubility of gas in the oil and other factors may have undetermined effects on thermogenic gas hydrate formation and stability. Berecz and Balla-Achs (1983) state that the presence of an oil phase destabilizes gas hydrates formed in pipelines, but present no quantitative estimates of the degree of destabilization. A similar situation would be expected to exist in sediments.

Occurrence of Biogenic Gas Hydrates

Biogenic gas hydrates have been recovered or are inferred to exist in a variety of geological settings in the Gulf of Mexico including abyssal plain, intraslope basin, and submarine trough. Cores from DSDP Sites 90 and 91 probably contained gas hydrates in sediments from the Sigsbee Abyssal Plain, as did cores from Site 89 from the lower continental rise. Gas hydrates were recovered from the Orca Basin, a deep, hypersaline intraslope basin on DSDP Leg 96. Hydrates were recovered from the Mississippi Trough and from various intraslope basins along the Louisiana continental slope whose exact configurations have not been released (Brooks and Bryant, 1985a)

Source of Biogenic Gas

Formation of gas hydrates requires not only proper pressure and temperature conditions, but also pore fluids which are supersaturated in gas (Makogon, 1978; Krason and Ridley, 1985a, 1985b). Thermogenic gas hydrates can form if quantities of gas sufficient to supersaturate pore fluids migrate to the gas hydrate stability zone from deep source beds. Most biogenic methane is formed prior to burial to 1,000 m (Rice and Claypool, 1981). Biogenic gas can saturate pore fluids by either efficient in situ gas production or short distance migration, or both. Offshore areas under proper pressure - temperature conditions with efficient biogenic gas production or accumulation should contain hydrates.

Type of Organic Matter

Very little information concerning types of sedimentary organic matter in the Gulf of Mexico is in the public domain. A mixture of terrestrial and marine organic matter would be expected to be deposited in the sea floor sediments. Because of the generally oxic conditions of the Gulf of Mexico, the majority of the marine organic matter present would be expected to be aerobically oxidized except where sedimentation was extremely rapid or in restricted intraslope basins where anoxic conditions exist, such as the Orca Basin.

Preservation of Organic Matter

Previous work has shown that a high sedimentation rate is an important factor favoring preservation of organic material (Muller and Suss, 1979; Ibach, 1982; Krason and Ridley, 1985a, 1985b). A high sedimentation rate minimizes the time that the sedimentary column is exposed to aerobic oxidation, thereby preserving a greater portion of the original sedimentary organic matter for subsequent anaerobic methanogenesis. Sedimentation rates are generally reported in units of cm/1,000 y. Compaction with burial results in sediment accumulation rates appearing to diminish with depth. To compensate for compaction and to make possible mass comparisons between organic and inorganic sediments, the downward flux of sediment through the section is calculated. Sediment flux is obtained by the following equation:

$$\text{Flux (mg/cm}^3\text{yr)} = SR \times (1 - \emptyset) \times \rho.$$

where SR = Sedimentation rate
$\emptyset$ = Porosity
ρ = Grain density (2.65 g/cm^3)

Sediment flux as a function of depth are shown in Figures 50 and 51 for three DSDP sites where hydrates were suggested (89, 90, and 91) and for Site 3 where no hydrates were indicated. No obvious relationship between sediment and organic matter input, and gas hydrate presence or absence can be discerned from these plots. The direct relationship of clastic sediment flux and organic matter flux indicates oxic conditions were dominant during sedimentation at all of the sites illustrated.

Sparse data exist on the oxygenation of bottom waters in the Gulf of Mexico. A consensus exists that the region is generally well oxygenated. McKee et al. (1978) presented data showing bottom water to contain 4 to 5 ppt of oxygen. If such values are typical, sedimentation rate and sedimentary

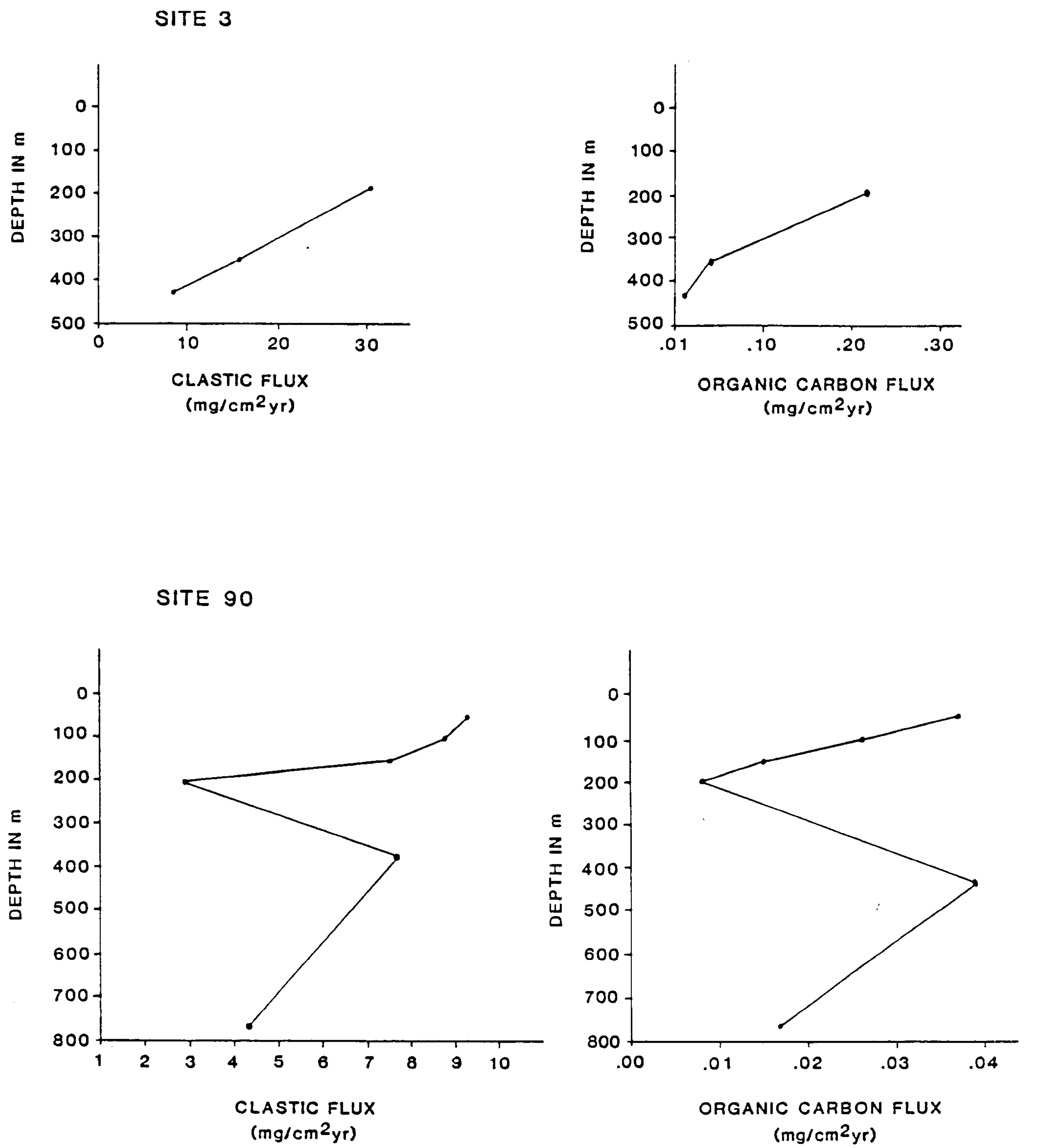


Figure 50. SEDIMENT FLUX, DSDP SITES 3 AND 90

After Worzel et al. (1973)

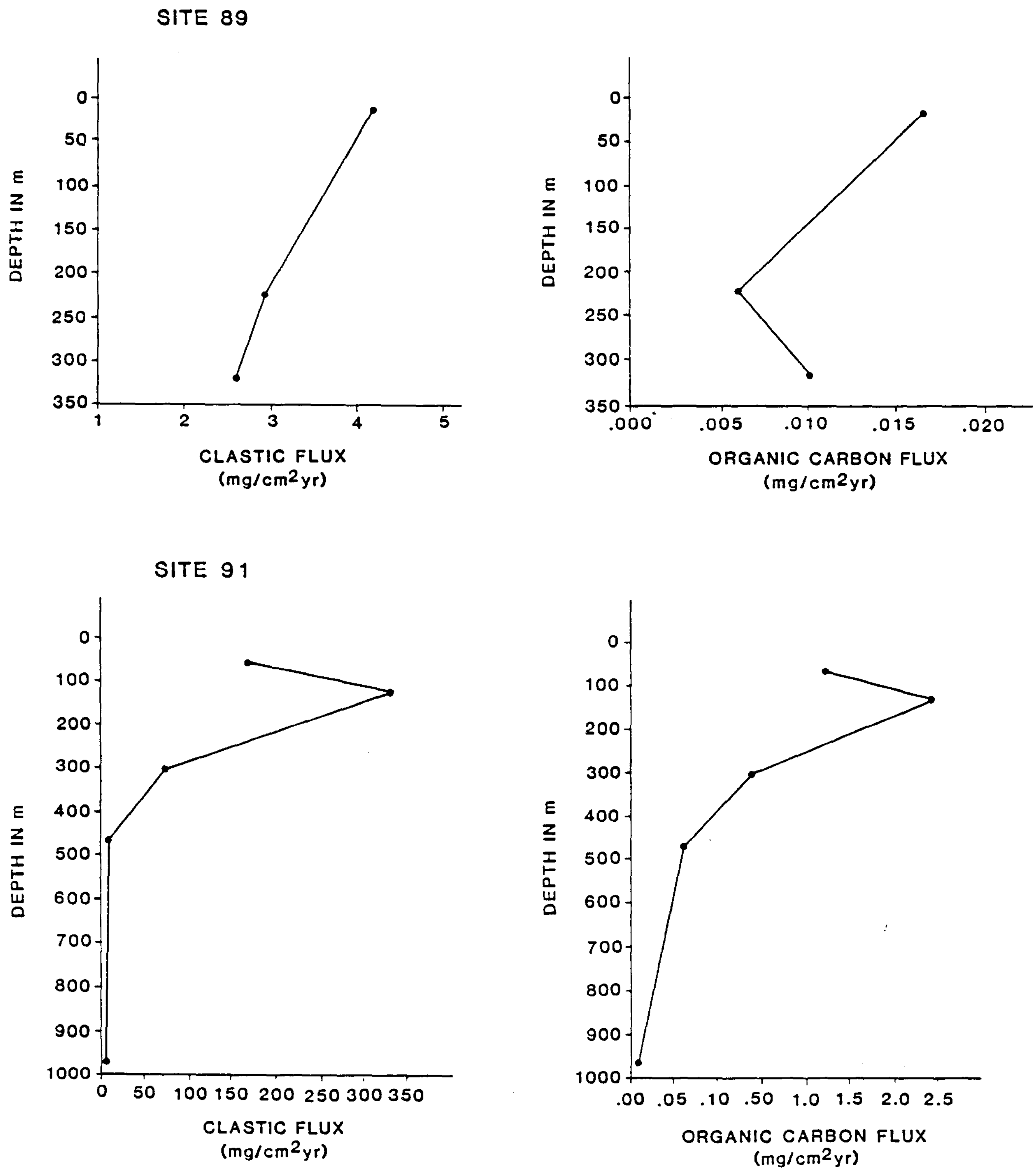


Figure 51. SEDIMENT FLUX, DSDP SITES 89 AND 91

After Worzel et al. (1973)

organic content are probably controlling factors for gas formation. Data available from NOAA (1974) indicates that throughout the western Gulf region an oxygen minimum of 2.6 ppm occurs at 400 to 500 m with oxygen concentration increasing to 4.7 ppm at 1,500 m, the greatest depth reported (Figure 52). These values agree with previously stated contentions that rapid sedimentation, very high organic productivity, or large input of allochthonous detrital organic matter is necessary for preservation of organic material through the zones of aerobic and sulfate reducing diagenesis in the Gulf of Mexico.

The Orca Basin is a major exception to the generally oxic depositional environment of the Gulf of Mexico. The Orca Basin is an intraslope basin located in 2,400 m of water with an area of about 240 km^2. The floor of the basin is covered by a layer of brine 200 m thick with salinity of 24%, eight times that of seawater. The oxygen content of the brine drops rapidly to near to zero (Trabaut and Presley, 1978) due to the density of the brine (1.19 g/cm) inhibiting downward oxygen transport.

Unlike DSDP Sites 88, 89 and 92, the salinity of the pore water in the Orca Basin diminishes with depth, indicating that the source of the brine is not an underlying diapir (Sackett and Bernard, 1977). The apparent source of salt for the brine is one of the flanking diapirs which coalesced to form the basin being exposed to the water by mass wasting of sediments.

The brine has produced an unusual biological environment. The unusual salinity seems to inhibit sulfate reducing bacteria; sulfate is depleted with depth in the sediments but at a much slower rate than typical, requiring 5.5 m to reach 0.7 ppm (Sackett and Bernard, 1977). Weisenburg (1979) noted that the brine of the Orca Basin contains the highest concentration of methane ever measured in seawater (880 - 1070 mL/L). The C values of the brine and sediment indicate that methane is being consumed in the near-surface sediments; but, due to high methane productivity in deeper sediments, some methane escapes into the brine. Since the density stratification of the brine precludes convective mixing, the methane can apparently escape only by diffusion, a slow process resulting in high levels of methane in the brine.

Biogenic gas hydrates were recovered in DSDP Leg 96 from the Orca Basin (Brooks and Bryant, 1985a). The biogenic hydrates ranging up to a few centimeters in size were recovered at a subbottom depth range of 25 to 47 m. Biogenic gas was encountered in abundance throughout drilling to a total depth of 92.5 m. The depth of occurrence of the hydrates coincides with black organic and/or pyrite-rich muds. Shallower than approximately 20 m, oxidized Holocene slump deposits dominate; at depths greater than about 50 m gray, Wisconsin age muds dominate. Results of geochemical analyses of hydrates, sediments, and pore waters are not available as this report is prepared.

Total Organic Carbon Content

Although there is no direct relationship between total organic carbon (TOC) of a sediment and biogenic methane productivity, some minimal amount of organic material must be present to sustain bacterial methanogenesis. One minimum necessary TOC estimate often quoted is 0.5% (Rice and Claypool, 1981).

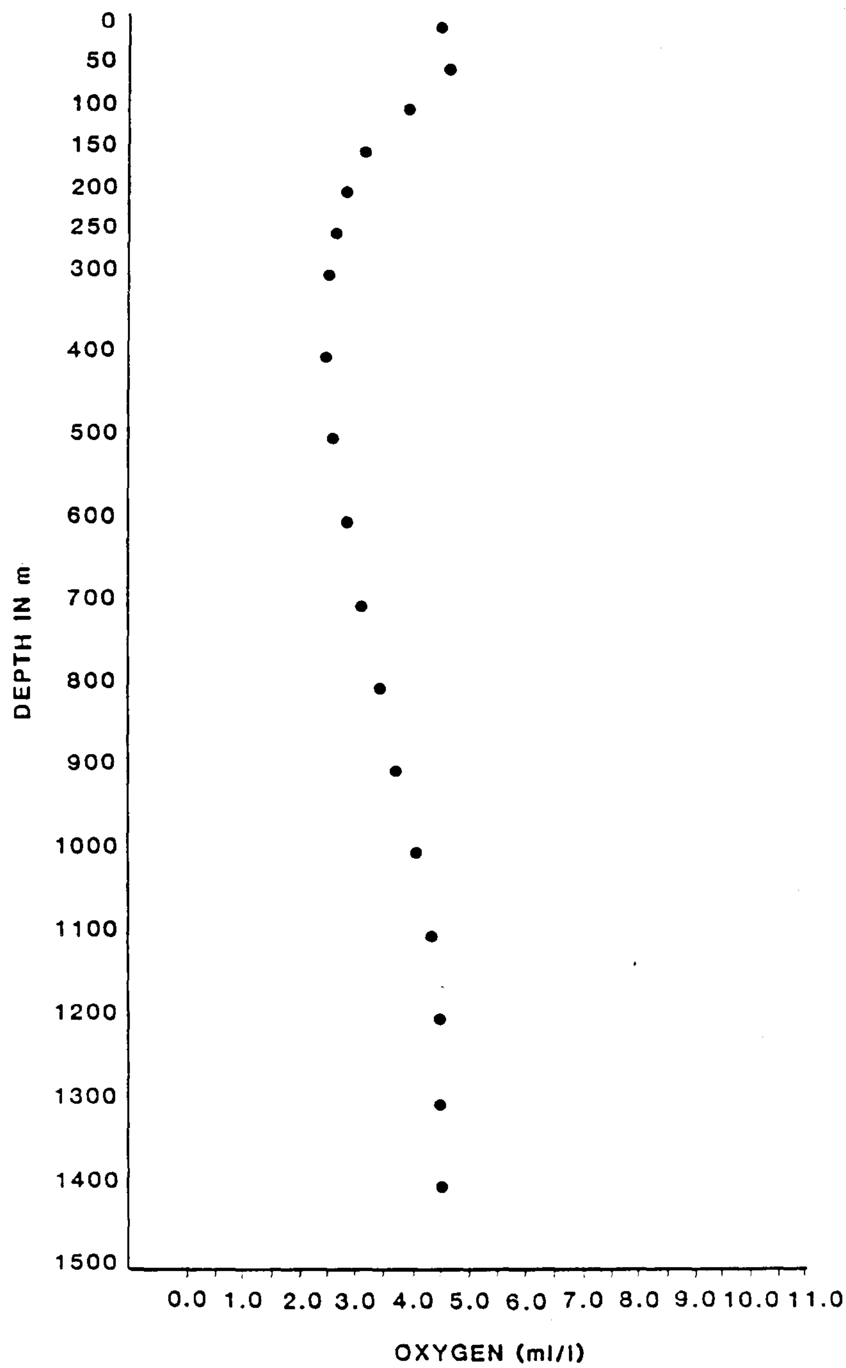


Figure 52. MEAN OXYGEN PROFILE FOR GULF OF MEXICO

After NOAA (1974)

Although thousands of sediment cores have been taken from the Gulf of Mexico, few besides DSDP cores have been analyzed for organic carbon content. The U.S. Navy collected cores of Gulf of Mexico sediments during the 1968 cruise of the USS San Pedro and tested for TOC. The locations and mean values for the upper 3 meters of sediment are plotted in Figure 53 along with corresponding values from nearby DSDP sites for comparison. The TOC values for the DSDP holes are substantially lower than those collected by the Navy. The USS San Pedro data set comprises two cruises, the March and the May cruises. Some difference exists between the March and May cruises. A statistical procedure, one way analysis of variance, can determine whether differences between groups of measurements are so great that inaccurate analytical determinations are likely. A one way analysis of variance indicates that a significant difference exists between the DSDP results and the Navy results at a confidence level of 99%. A test for difference between the two sets of Navy data indicates that the results are significantly different at a confidence level of 94%. Boyce (1973) of the DSDP team used the LECO furnace method. That method has a relatively low precision (<0.06% absolute). However the method is thought to be fairly accurate. The LECO furnace technique is the defacto standard method for TOC determination in the geochemical community. Thus high confidence is placed on the DSDP organic carbon values. The Navy scientists used the Allison method which may not have the accuracy of the LECO method.

One estimate of TOC values for sediments from offshore Louisiana and Texas was made by Booth (1980) from 34 m cores from 15 areas on the outer continental shelf and continental slope. Booth obtained a mean TOC value of 0.93% from 305 samples. Values ranged from 0.21% to 3.73%.

Tests were made on cores recovered by DSDP Leg 10 for total organic content; the results of which are presented in Table 5.

Of the four sites from which gas hydrates are considered to have been drilled, only Sites 90 and 91 can confidently be examined for correlation of biogenic gas productivity and organic carbon. Site 88 was shown by chromatographic and isotope analysis to probably contain some migrated thermogenic gas; the organic content of the host sediments would not be expected to have any bearing on gas content under such conditions. Since no detailed chemical analyses were performed on gas from Site 89, its biogenic or thermogenic origin is unclear. Site 89 is located close to the Mexican Ridges fold belt in the Bay of Campeche where much structural disturbance in the form of shale diapirism, thrust faults, and growth faults has been documented (Buffler et al., 1979). These features may have permitted migration of deep thermogenic gas to levels drilled at Site 89. Sites 90 and 91 contained proven biogenic gas from undisturbed abyssal sediments.

Sediments from Sites 90 and 91 provide evidence that sediments considered poor in organic content may have efficient bacterial methanogenesis. Gas was noted at depths of 130 to 767 meters at Site 90 with the greatest amounts of methane being evolved at 272 to 348 m (Worzel et al., 1973). The mean organic carbon content for the productive interval was 0.33% (Table 5). The two samples tested from the interval with maximum gas content contained 0.5% TOC. At Site 91, vigorous degassing was noted from 159 to 838 m subbottom depth. The mean TOC of samples from this depth interval is 0.45%. These determinations were made on small samples of sediment. Inhomogeneity in the sample stack could have resulted in skewing

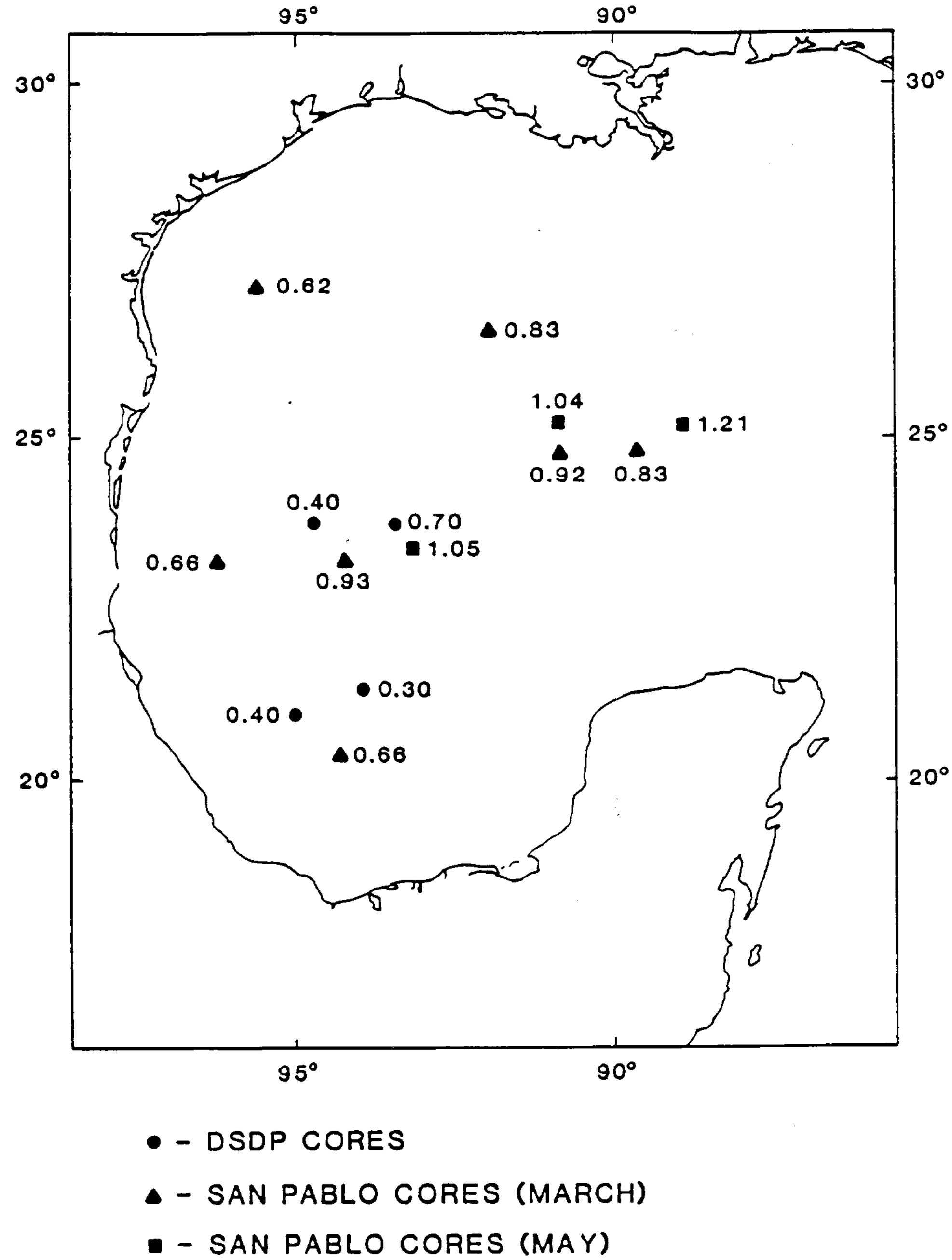


Figure 53. ORGANIC CARBON CONTENT OF GULF OF MEXICO CORES, in percent total organic carbon (TOC). Average of top 20m of sediment.

TABLE 5.

ORGANIC CARBON AND CARBONATE CONTENT OF SEDIMENTS FROM DSDP LEG 10

Site	Core, Section	Depth (m)	Organic Carbon (%)	$CaCO_3$ (%)
88	1-1	0.15	0.3	34
	2-1	51.15	0.4	29
	4-1	104.10	0.3	35
89	1-1	1.31	0.4	33
	4-1	220.17	0.2	50
	6-3	379.16	0.4	19
90	1-1	0.09	0.4	18
	2-1	70.70	0.3	41
	2-2	71.60	0.3	25
	3-2	131.65	0.2	42
	4-2	189.65	0.3	26
	5-1	236.10	0.5	10
	6-1	293.22	0.5	10
	11-6	682.65	0.4	19
	13-3	766.42	0.1	5
91	1-5	66.20	0.6	22
	2-2	124.70	0.7	13
	2-3	126.20	0.5	16
	3-1	159.20	0.7	18
	4-1	177.00	0.7	22
	5-1	186.10	0.4	18
	5-5	192.10	0.6	
	6-1	301.10	0.3	
	12-3	782.40	0.4	
	14-2	800.40	0.4	
	15-1	807.10	0.3	
	16-1	816.10	0.3	

the mean downward; thin, organically rich horizons which were not sampled may be the sources of the abundant methane. However, the large numbers of samples (n = 15) would indicate that these lower than expected TOC values are probably valid.

Vertical Distribution of Biogenic Gas Hydrates

The vertical distribution of biogenic gas hydrates is dependent on pressure, temperature, and microbial activity. Gas hydrate formation requires saturation of pore spaces with methane (Makogan, 1978). Maximum propensity for gas hydrate development would thus be expected where methane production is the greatest. Information on the optimum temperature (depth) for bacterial methanogenesis is scarce; Rice and Claypool (1981) state that most biogenic methane is produced prior to burial to 1,000 m. The depth of maximum methanogenesis is probably variable, dependent on quantity and type of substrates, presence of trace nutrients, salinity, etc.

Because of in situ methane generation, the deepest occurrence of biogenic methane may be considerably shallower than the limit of equilibrium pressure and temperature conditions of gas hydrate stability. However, it could also be argued that because most biochemical processes such as bacterial metabolism approximately double for each 10°C rise in temperature (Lehninger, 1975), biogenic methane production would be maximized in the warmer sediments beneath the base of the gas hydrate zone (approximately 23°C) and above the level at which geothermal heat hinders the microbes' biochemical processes. In that case, the methane supply for gas hydrates would be produced from below and migrate to the gas hydrate stability zone, similar to the situation for thermogenic gas hydrates. With biogenic gas being supplied from below the hydrate zone, the deepest occurrence of gas hydrates would be expected to coincide with the highest possible temperature of gas hydrate formation under prevailing pressure and gas saturation conditions.

Since gas hydrates were not expected on DSDP Leg 10, no provision for determining the quantity of dissociated gas evolved from the cores was on board. Subjective descriptions by Worzel and Bryant (1973) of which cores degassed most violently do provide some insight into the depth distribution of gas hydrates (Figure 54). None of the sites from which biogenic hydrates are likely to have been drilled (89, 90, 91) had anomalously gassy cores from depths less than 100 m. Site 89 had a progressive increase in gassiness from 119 m to 380 m whereupon it decreased rapidly to a total depth of 440 m. Cores from Site 90 increased in violent degassing from 130 m to a maximum at 237 to 348 m and tapered off to the total depth of 767 m. Cores from Site 91 from 159 to 838 m degassed but to a much lesser degree than cores from Sites 88, 89, and 90. At Site 91, no degassing was mentioned from 839 m to a total depth of 900 m. This may indicate that bacterial methanogenesis ceased below 840 m due to consumption of available nutrients by the organisms. An alternate explanation of the observed gas distribution is that only gas in hydrate form would be released from the sediment slowly enough to display the reported violent degassing upon recovery. Thus, free interstitial gas may have been present below 840 m at Site 91, but the nonhydrated gas may have escaped before the cores were brought on deck. If that was the case, 839m could represent the critical pressure-temperature conditions for hydrate stabilization in the abyssal plain.

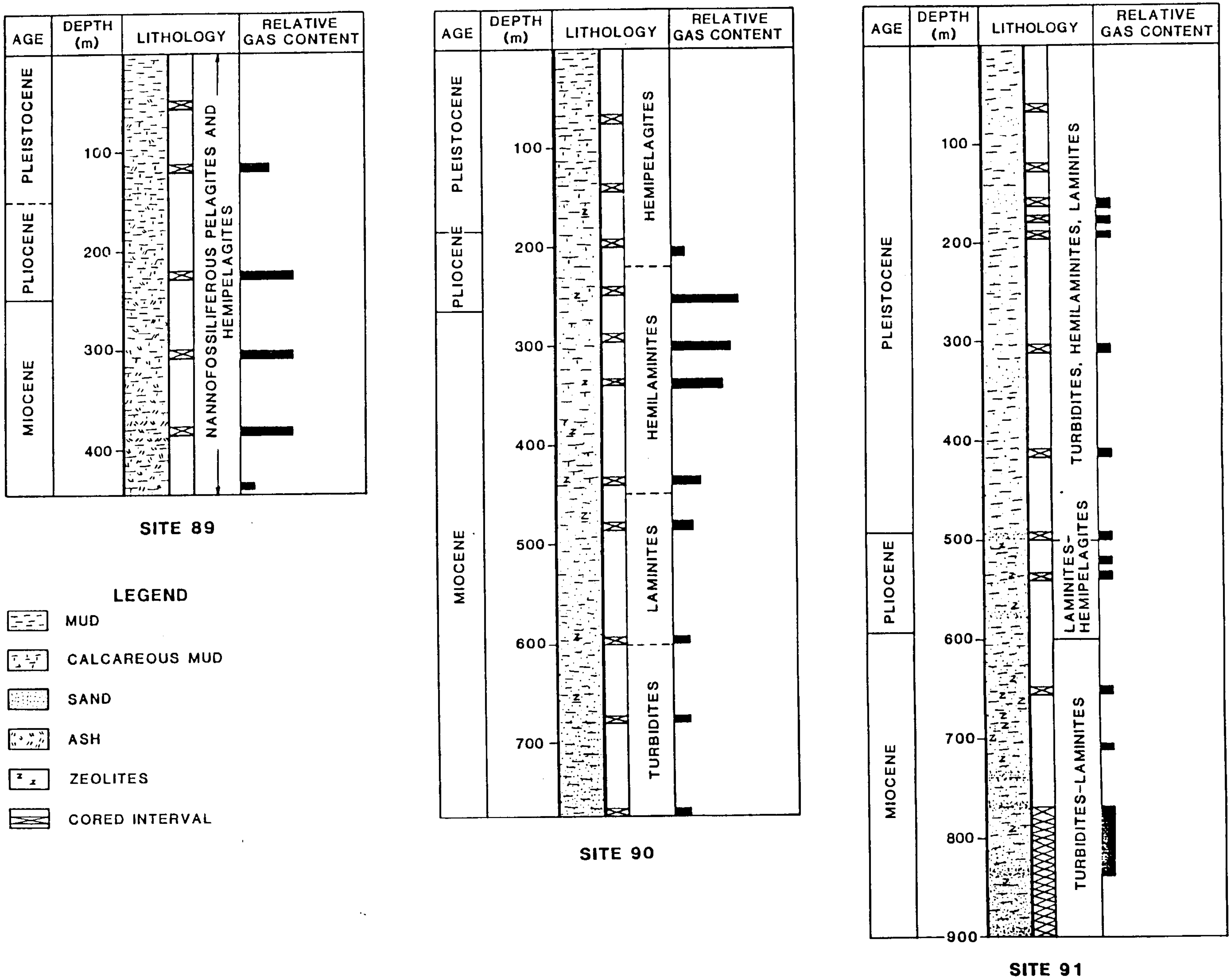


Figure 54. LITHOLOGY AND GAS CONTENT OF SEDIMENTS FROM DSDP SITES 89, 90, 91

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Gas Hydrate Host Sediments

Sediments in which gas hydrates are found may have a bearing on formation and stability factors. Gas hydrate host sediments can be considered from the standpoints of both gas hydrate matrix media and as a source of the gas. Examination of the sediments as sources was inconclusive; sites at which gas hydrates were found had relatively low TOC values. Evaluation of the host sediments in terms of their relationship of lithology to gas hydrate stability follows.

No consistent lithological distinction could be made between sediments above and below the shallowest occurrence of gas hydrates on DSDP Leg 10 (Figure 54). The upper 100+ m of Sites 89, 90, and 91 did not display the violent degassing associated with gas hydrates in cores. Much shallower occurrences of gas hydrates in sediments of the northern Gulf of Mexico have been documented (Brooks and Bryant, 1985a). It is possible that some disseminated hydrate may exist at similarly shallow subbottom depths at these DSDP sites, but we interpret the data to indicate that a confident inference of gas hydrate presence can be made only for cores deeper than 100 m. At Site 89 the transition to gas hydrates containing sediments between cores 2 and 3 is marked only by a slight increase in ash content. Above and below the first appearance of probable gas hydrates, the sediments are similar greenish gray nannofossiliferous ooze. The appearance of probable gas hydrates at Site 90 is marked by a slight change from a clayey silt with 30% silt size grains to a clayey ooze with 13% silt. Zeolites, presumably of volcanic origin, are also found in the cores beneath the first appearance of probable gas hydrates. The appearance of probable gas hydrates at Site 91 at 159 m is not marked by any change in lithology; gray silty clay laminites to turbidites predominate throughout.

The zone of maximum core degassing at Site 90 from 237 to 348 m corresponds with a late Miocene zeolitic hemilaminite section in a transition from deeper turbidites to shallow pelagic deposits. The silt content of 18% to 26% reflects the hemipelagic character of the sediments.

No maximum degassing zone was noted for Site 91; the entire cored section from 160 to 838 m degassed, although less vigorously than at Sites 88, 89, and 90. The gassy middle Pleistocene through middle Miocene section is principally composed of laminites and turbidites. Occasional zeolites and ash deposits were noted. No distinct lithological difference exists between the gassy cores and deeper cores (840 - 900 m) which did not degas noticeably.

Suspected gas hydrate degassing at Site 89 reached a maximum at 380 m. The probable hydrates were found both above and beneath a major lithological break at 350 m from an upper ashy carbonate ooze to an ashy hemilaminite. The degassing decreased rapidly from the maximum in the hemilaminite sediments in the 20 m to total depth. No cores were pulled, and no record of cuttings descriptions was published from this zone of diminishing gassiness; thus, no lithological interpretation of it can be made.

Pore Water Salinity

Analysis of gas hydrate bearing sediments cored from the continental slope offshore of Guatamala revealed that pore water decreased in salinity with depth in the presence of gas hydrates (Hesse and Harrison, 1981).

Pore water analyses by Manheim et al. (1973) from DSDP Sites 89, 90, and 91 do not show the same decrease in salinity with depth (Table 6). The interstitial water from Site 89 nearly doubles in salinity by 224 m depth, the depth at which maximum degassing was noted (Worzel and Bryant, 1973). Site 89 displays a slight shift in salinity from 35.2 to 32.4 ppt between 7 and 73 m depth; however, sulfate is depleted from 2.39 to 0.22 in the same interval suggesting that the observed salinity change is due to sulfate reduction. Sulfate reduction can decrease salinity of pore water by precipitation of sulfide product ion by iron, removing the charged species from the solution. No significant decrease in salinity was reported between a gas free core at 133 m to a gassy core from 238 m. Pore water from Site 91 increased in salinity in the transition to gassy cores between 127 and 167.

Seismic Evidence

BSRs have been reported from the Gulf of Mexico by Shipley et al. (1979), Buffler et al. (1979), Hedberg (1980), Buffler (1983), and Brooks and Bryant (1985a).

In an article on generation and migration of methane in sediments, Hedberg (1980) presented two seismic sections from the Gulf of Mexico showing distinct BSRs. Since the profiles were provided by Gulf Oil Corporation, the exact locations of the sections are proprietary and were not included in the article. The two sections are reproduced here as Figures 35 and 38. In Figure 35, the BSR is a distinct reflector between 3.25 and 3.50 seconds travel time. The reflector is discordant with bedding, which enhances its visibility. The BSR disappears beneath the syncline in the right center of the figure. It is not readily apparent whether the hydrate layer itself is absent in the syncline, or if the lack of pooled gas beneath a gas hydrate cap in the structural low suppresses the BSR. The BSR maintains a constant depth of about 0.50 sec subbottom in water depths of 1,900 to 2,300 m. Figure 38 illustrates another BSR which is similarly limited to an anticline which overlies a shale diapir. The reflector's identity as a gas hydrate horizon is substantiated by its discordant relationship to nearby sedimentary layers. The BSR does not parallel the sea floor exactly; it is deeper at the crest of the anticline (0.54 sec) than on the flank (0.48 sec).

Other literature citations of BSRs in the Gulf of Mexico (Buffler et al., 1979; Shipley, 1979; Buffler, 1983) describe anomalous reflectors on seismic sections collected by the University of Texas Marine Science Institute (since renamed the University of Texas Institute of Geophysics). In his paper on BSRs on continental slopes and rises, Shipley (1979) diagrammed the subbottom depth to ocean water depth relationship for six BSRs from the Gulf of Mexico, demonstrating that they were probably gas hydrate boundaries rather than diagenetic features. He referenced the seismic lines to an article on the structure of the southern Mexican Ridges fold belt by Buffler et al. (1979). In that article, six subparallel seismic lines from offshore of the Golden Lane area offshore of the Tampico Embayment between 21°N - 22°N and 95°W - 97°W were reproduced. Buffler et al. identified one BSR in a figure caption as a possible gas hydrate horizon. In the introduction to a subsequent presentation, Buffler (1983) stated that possible gas hydrate zones were present in anticlines on several of the lines without explicitly identifying which reflectors he had so interpreted.

TABLE 6.

PORE WATER SALINITY OF DSDP SITES 89, 90, AND 91

Site	Depth (m)	Salinity (ppt)	Hydrates Present*
89	3	32.3	
	121	41.1	*
	224	65.3	*
90	7	35.2	
	73	32.4	
	133	32.7	
	238	32.9	
91	65	32.01	
	127	32.2	
	167	34.12	*
	538	34.19	*

We have confirmed BSRs at at least 37 locations in the southwestern Gulf of Mexico. Single and multiple channel seismic reflection profiles covering 38,500 km were carefully examined for anomalous reflections. Bottom simulating reflectors were observed in seismic sections from the southern Mexican Ridges fold belt and the Campeche Knolls from the 1969 cruise of the US Navy research vessel Kane, the 1971 USGS - IDOE (International Decade of Ocean Exploration) Gulf of Campeche survey, and the 1975 Golden Lane and Western Gulf survey by the University of Texas Institute of Geophysics (UTIG). Locations of the BSRs are mapped on Figure 55. Due to the abundance of BSRs, only selected representative examples are discussed and illustrated in this report.

Although seismic records from throughout the Gulf of Mexico were examined for BSRs, strong BSRs were only recorded in the southern part of the study region, between 19° and 22° North latitude. The BSR locations can be divided into three groups by the subsea topography and underlying structures (Figure 56). The northwestern grouping occurs in the Golden Lane area of the main fold belt of southern Mexican Ridges between 20.6°N and 22°N. Here, the folds show abundant relief above the sea floor (Figure 57). South of this zone, many BSRs were found in the Bay of Campeche between 19.3°N and 20.5°N (Figure 56). In this area the sea floor exhibits much less relief as indicated by the regular distribution of isobaths in Figure 55. Seismic sections from this area reveal that large, northerly trending anticlinal structures similar to those farther north are present in the subsurface, but that their sea floor topographic expression has been greatly subdued by sediments ponded in adjacent synclines (Figure 58). A small number of BSRs were found in the Campeche Knolls salt dome province east of the two previously described areas (Figure 55). The anomalous reflections from this area were not as strong or as continuous as those from the anticlines to the west.

Southern Mexican Ridges

Seismic coverage of the southern Mexican Ridges area (Figure 56) includes single channel east to west lines 2-17, 2-20, 4-2, and 2-22 obtained by the U.S. Navy (Figure 59) and seven multichannel lines collected by UTIG. Between 21°N and 22°N the Mexican Ridges trend approximately N 30°W. Six of the UTIG lines -- GLR 16-1-7-9, GLG-18, GLG-22, GLG-20, GLG-24, GLR 16-1-10 -- are approximately normal to the ridges trending N60°E (Figure 55). The remaining UTIG line, WG-3, is oblique to the ridges and the other lines trending N60°W (Figure 59). Line WG-3 crosses all other seismic lines in the area.

Strong BSRs were recorded on two east to west single channel seismic lines 2-20 and 4-2 (Figures 57 and 60). More subdued BSRs can be traced on line 2-22 (Figure 61). BSRs on line 2-20 were previously reported by Brooks and Bryant (1985a). BSRs were also noted in five of the seven UTIG lines.

Sediment deposition concurrent with folding resulted in thick accumulations in the synclines and thinner sediment sections draped over adjoining anticlines. This depositional style resulted in sediment layers that are parallel to the sea floor only at the crests and troughs of the structures; the oblique relationship of sea floor and underlying sediments on the flanks of the structures permits detection of bottom simulating reflectors. Three seismic sections from the USNS Kane (2-20, 4-2, and 2-22; Figures 57, 60, 61) present

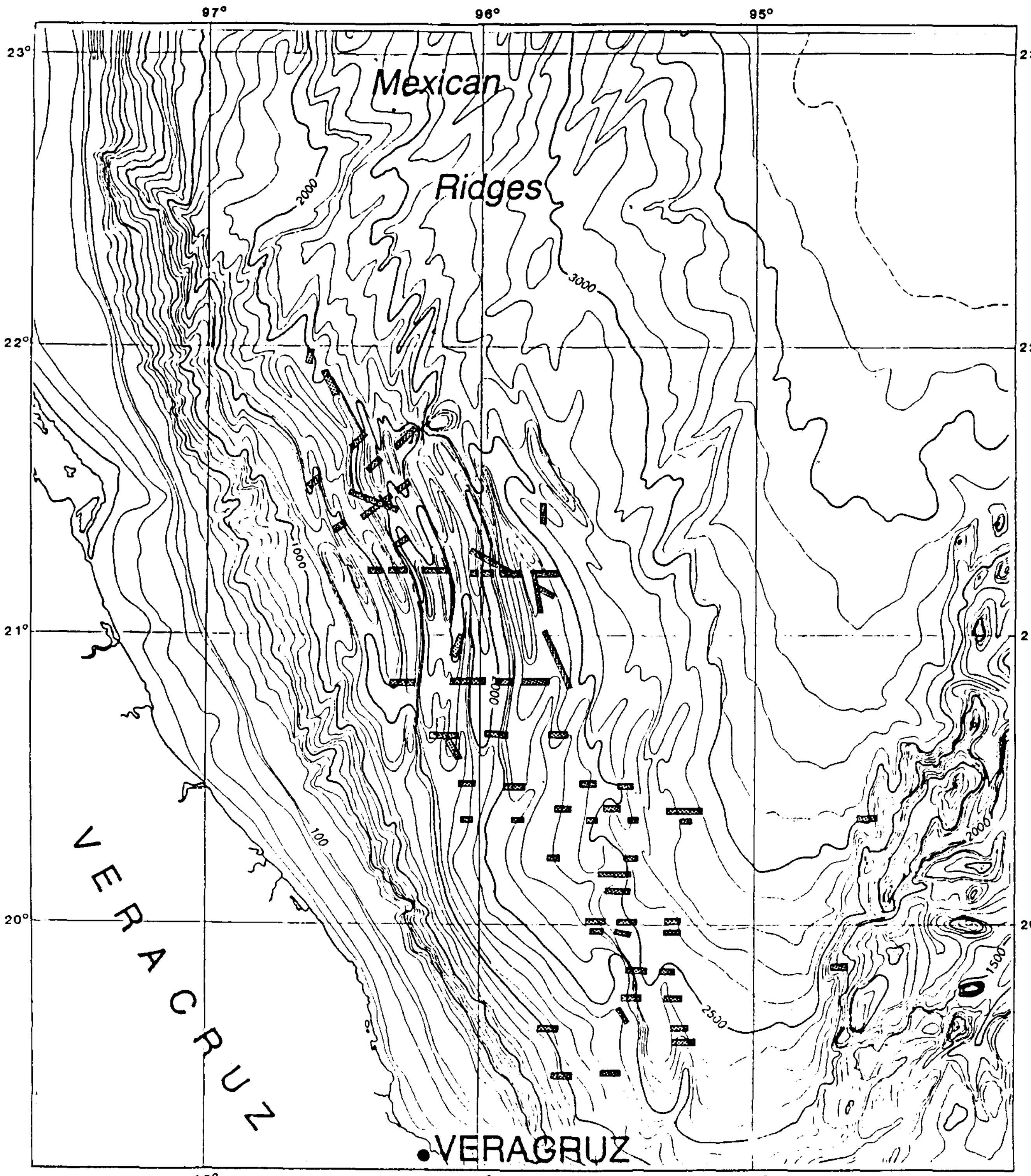
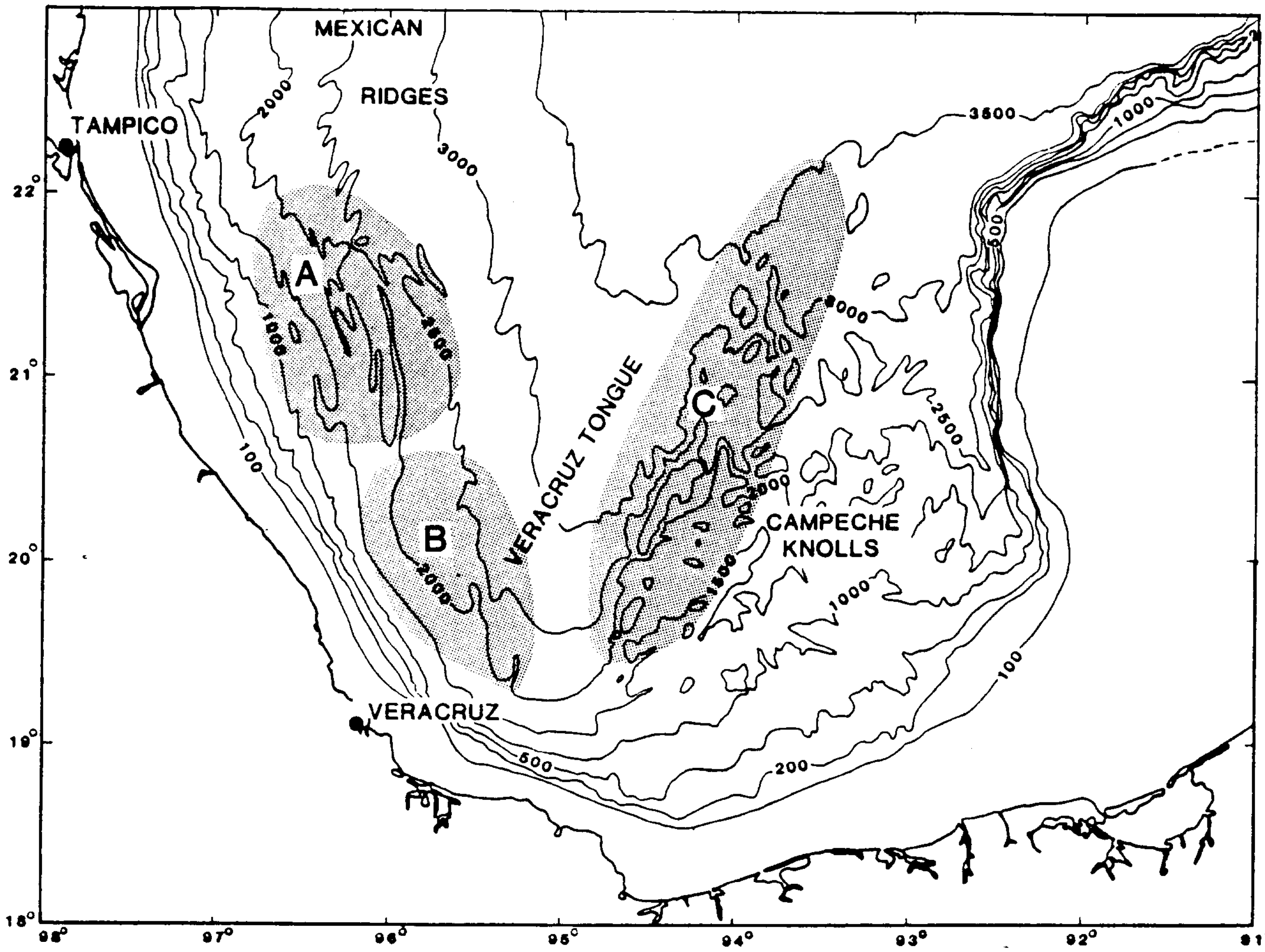


Figure 55. MAP SHOWING LOCATIONS OF BSRs IN THE WESTERN GULF OF MEXICO

Bathymetry from Bryant (1984)



A. Southern Mexican Ridges
B. Veracruz Continental Slope
C. Campeche Knolls

Figure 56. MAP SHOWING AREAS OF BSR OCCURRENCE

Figure 57, Seismic profile 2-20, southern Mexican Ridges, is located in the pocket at the end of the report.

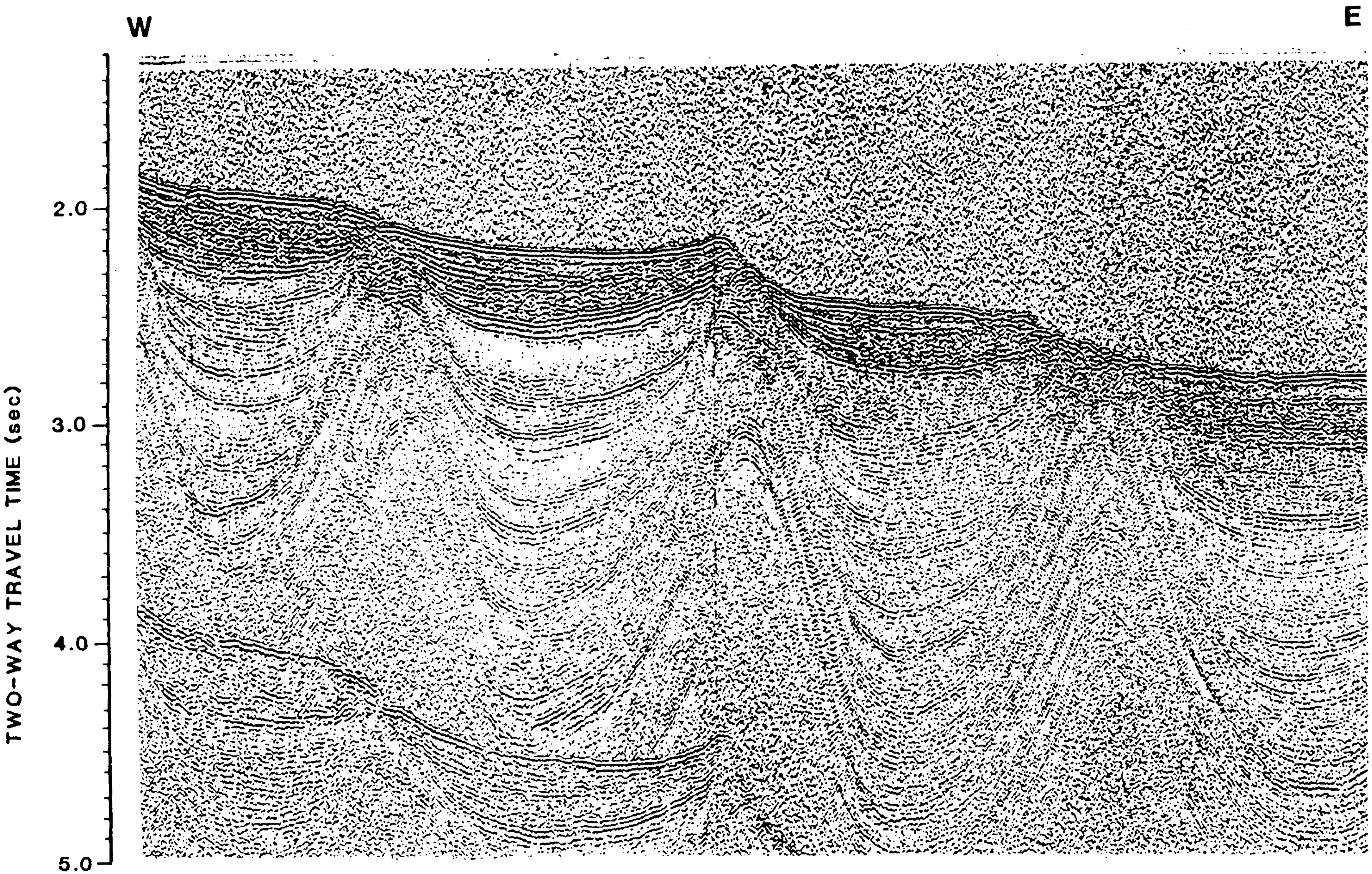


Figure 58 **SEISMIC PROFILE N, OFFSHORE OF VERACRUZ**

Structural style is typical of continental slope west of Campeche Tongue

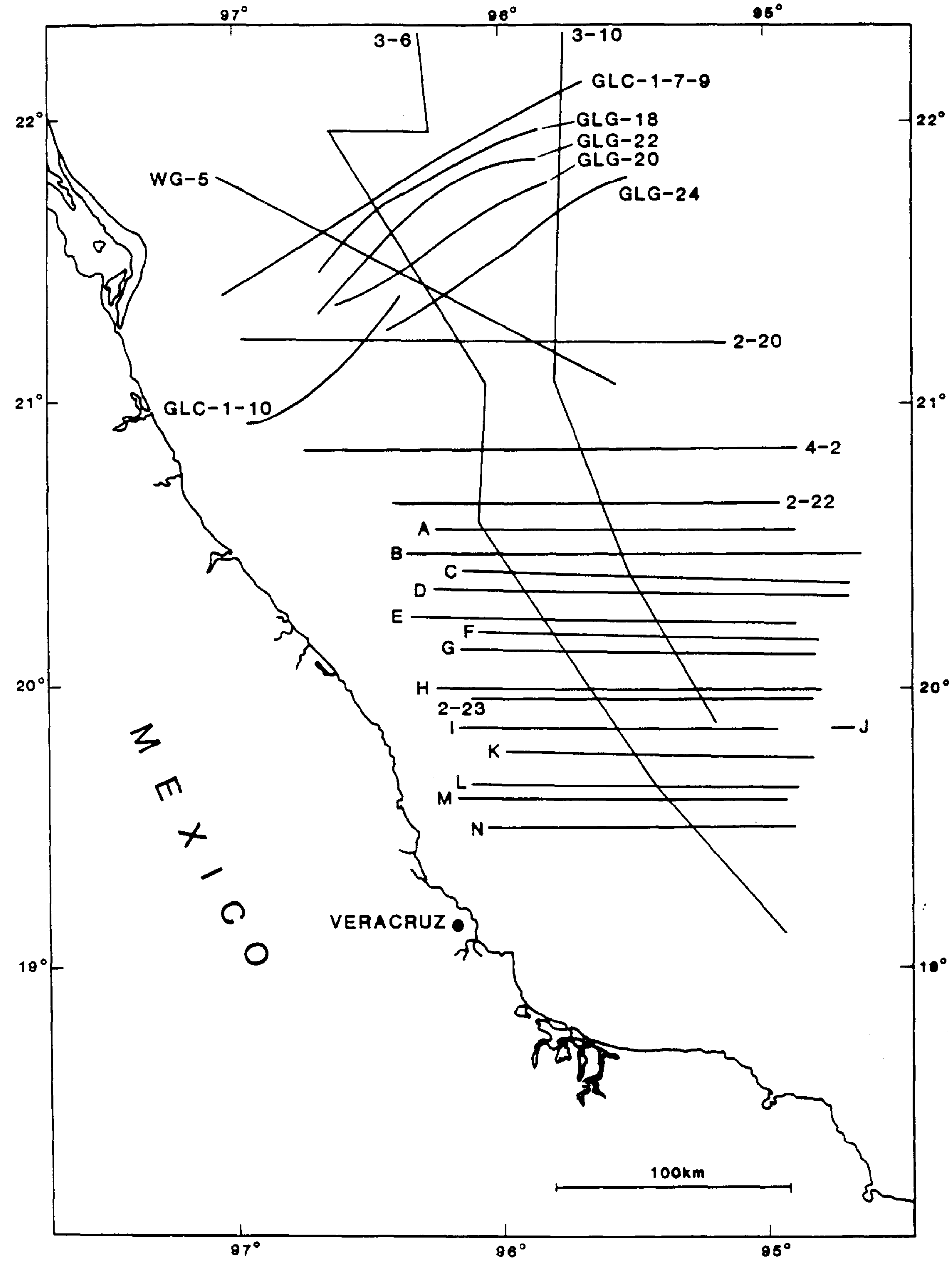


Figure 59. TRACK LINES OF SEISMIC PROFILES DISCUSSED IN TEXT

FIGURE 60, Seismic profile 4-2, southern Mexican Ridges, is located in the pocket at the end of the report.

FIGURE 61, Seismic profile 2-22, southern Mexican Ridges, is located in the pocket at the end of the report.

representative east to west cross sections of the southern Mexican Ridges area, with section 2-20 being the northernmost location, and 2-22 the southernmost.

The ridges cannot easily be correlated with certainty either by comparison between sections or by using published bathymetric maps, because the ridges are discontinuous over the 40 km spacing of the lines. The oblique multichannel lines show that similar structure is prevalent in the northern reaches of the area (21.2°N to 22°N; Figures 34, 36, 62). The closer spacing of these lines (5 to 20 km apart) allows visual correlation of some of the more prominent ridges, although seldom can they be traced over more than 20 km.

One unusually continuous ridge can be traced between profiles. The ridge which is well defined by the 2,000 m isobath in Figure 55, is presented as a series of cross sections in Figure 63. This ridge has a distinct BSR on both flanks in sections WG-5, 2-20 and 4-2 and a less well defined BSR on section 2-22 where the ridge is more subdued.

If it is assumed that the BSRs are continuous along the 100 km length of this ridge, an areal estimate of BSR extent on the ridge can be calculated. The BSR swells laterally (east to west) from about 2 km in the south (line 2-22) to 8 km in the central section (line 2-20). Assuming a roughly elliptical outline of the BSR in map view, the areal extent is approximately 450 km^2 for the BSR under the correlated ridge. Structural closure beneath the BSR varies markedly along the ridge from 60 m (0.2 sec) to as much as 840 m (0.7 sec).

One unusual aspect of the BSRs from this region is that they occasionally appear as a parallel pair of reflectors separated by approximately 0.1 sec two-way travel time (Figure 64). The doublet BSRs are only located in the two southernmost sections, 4-2 and 2-22. Multiple BSRs have previously been reported from the Gulf of Oman (White and Louden, 1983).

The seismic sections discussed above extend eastward into the Veracruz Tongue. The Veracruz Tongue east of the Mexican Ridges and west of the Campeche Knolls is a flat plain under 3,000 to 3,400 m of water (Figure 1). The sediments, presumably hemipelagic muds and oozes with possible turbidites, generally parallel the sea floor making detection of BSRs difficult. Distinct BSRs were not found in the deep waters. Some persistent reflectors in the depths expected for BSRs in this area (0.4 to 0.6 sec) were noted, but could not be confidently differentiated from sedimentary horizons (Figure 60).

One seismic cross section (4-2) does pass within 5 km of DSDP Site 89 where core degassing characteristics associated with gas hydrates were reported. To illustrate the relationship of seismic response to lithology and gas content, a comparison of section 4-2 and the DSDP cores is presented in Figure 65. Correlation is based on core data reported by Worzel et al. (1973) and seismic stratigraphy of Shaub et al. (1984).

In addition to the east to west profiles discussed, two north to south lines were shot on the Kane cruise. These lines intersect the east to west lines at 6 points allowing verification of BSR presence and an estimate of the continuity of the reflectors along the strike of the ridges.

Section 3-10 intersects lines WG-5, 2-20 and 2-22 at points where BSRs were recognized, and intersects line 4-2 at a point 5 km east of a BSR. There is a strong reflector on line 4-2 at the intersection with line 3-10 which has the appearance and subbottom depth expected of a BSR, but the sediments are parallel with the sea floor at that point and confident assignment of the reflector as a BSR could not be made.

FIGURE 62, Seismic profile WG-5, southern Mexican Ridges, is located in the pocket at the end of the report.

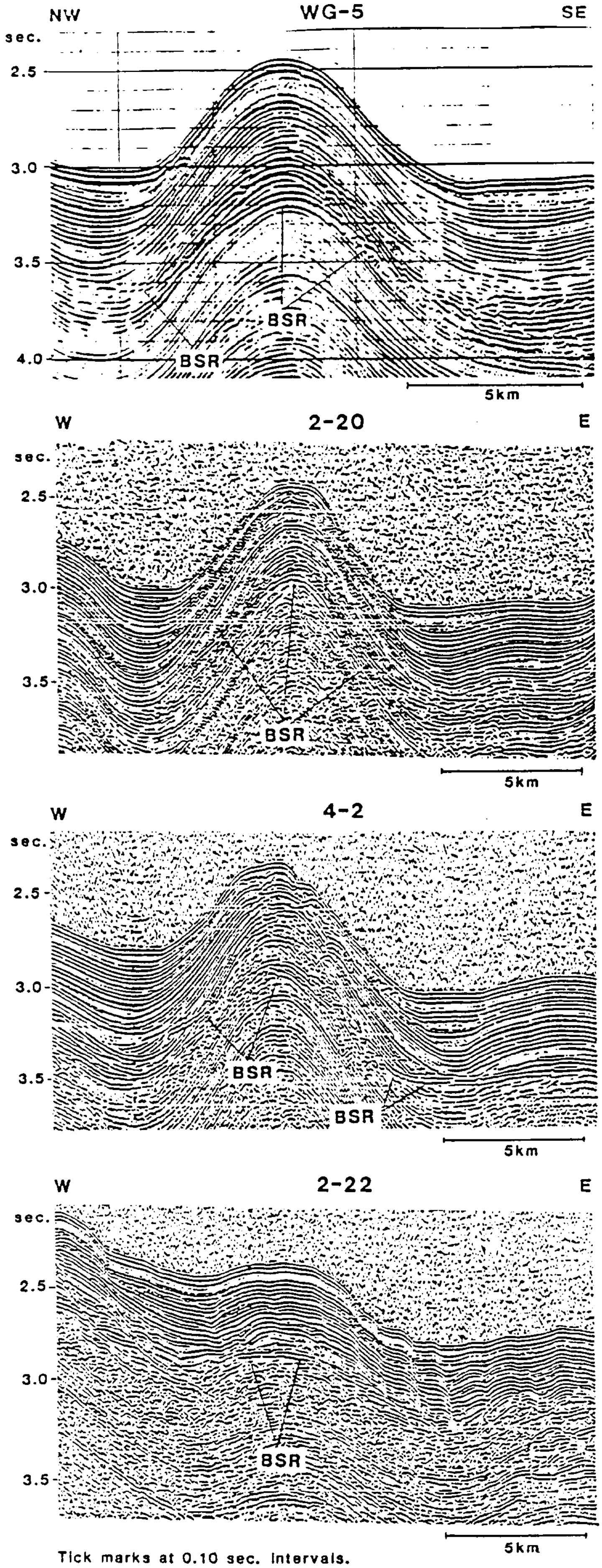


Figure 63. SEISMIC PROFILES OF ANTICLINE, SOUTHERN MEXICAN RIDGES

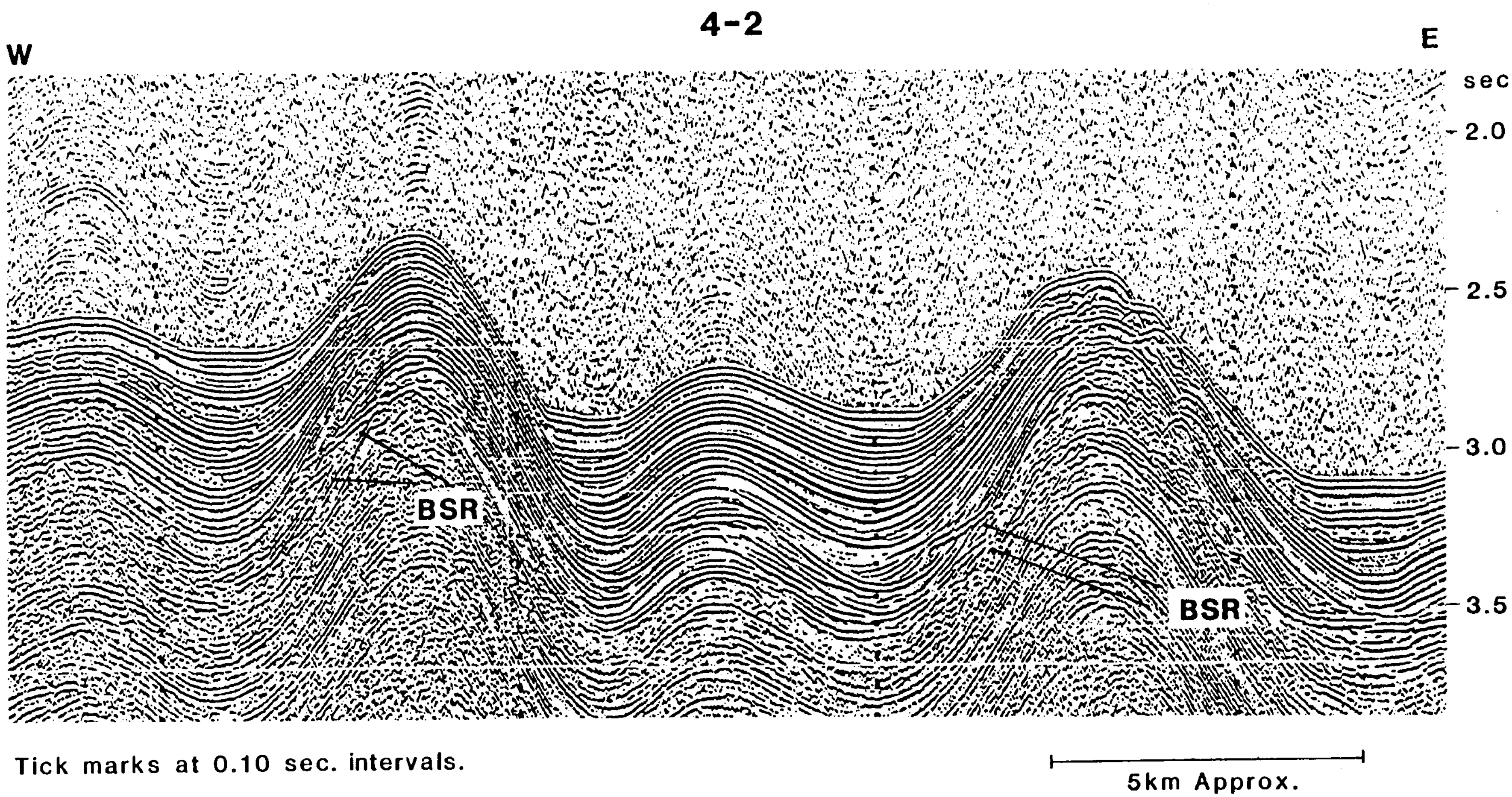
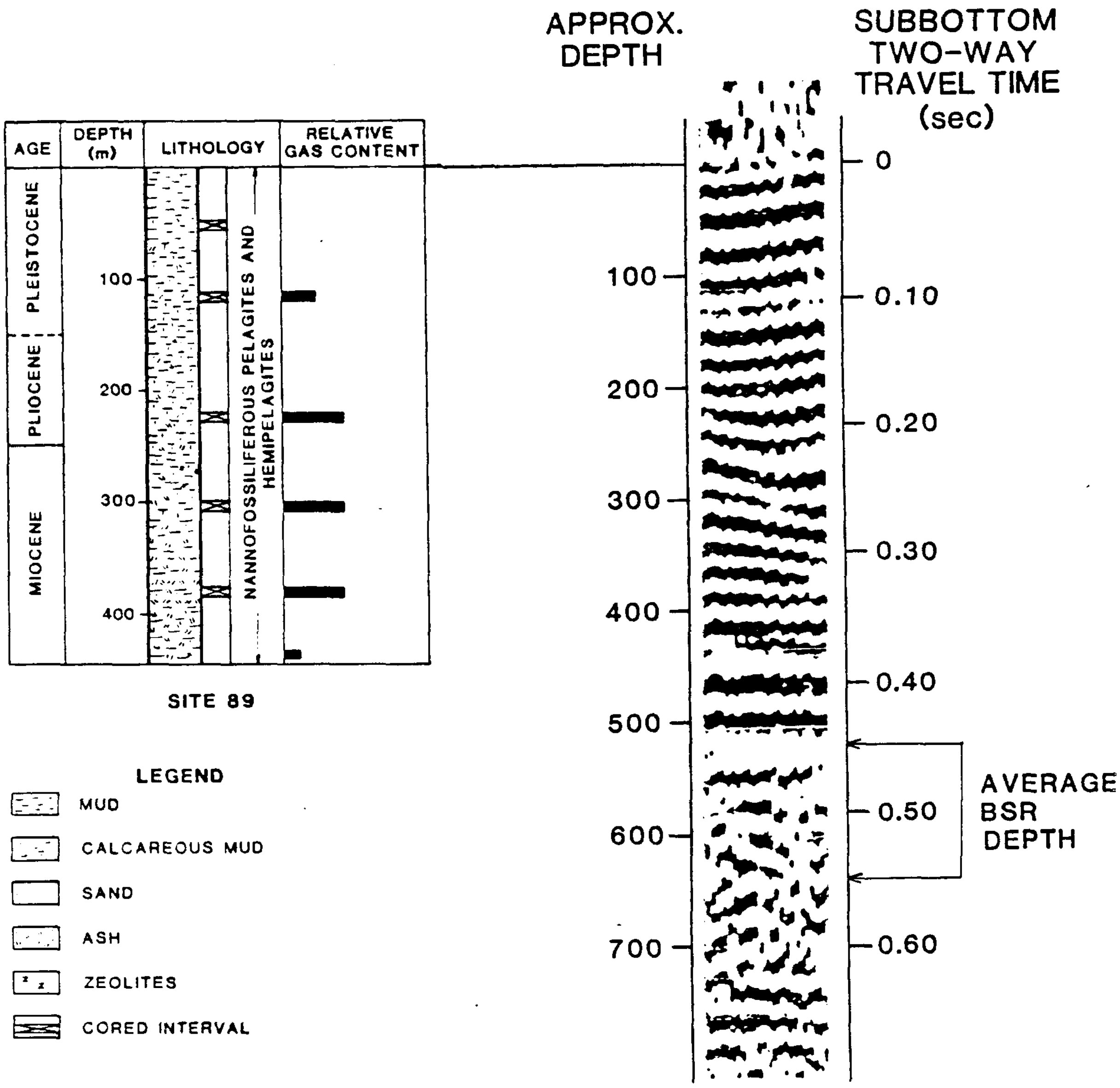


Figure 64. DOUBLE BSRs, SOUTHERN MEXICAN RIDGES



Approximate depth calculated assuming seismic velocity of 2300 km/sec.

Figure 65. COMPARISON OF DSDP SITE 89 CORES AND SEISMIC REFLECTIONS, LINE 4-2

Site 89 data from Worzel et al. (1973)

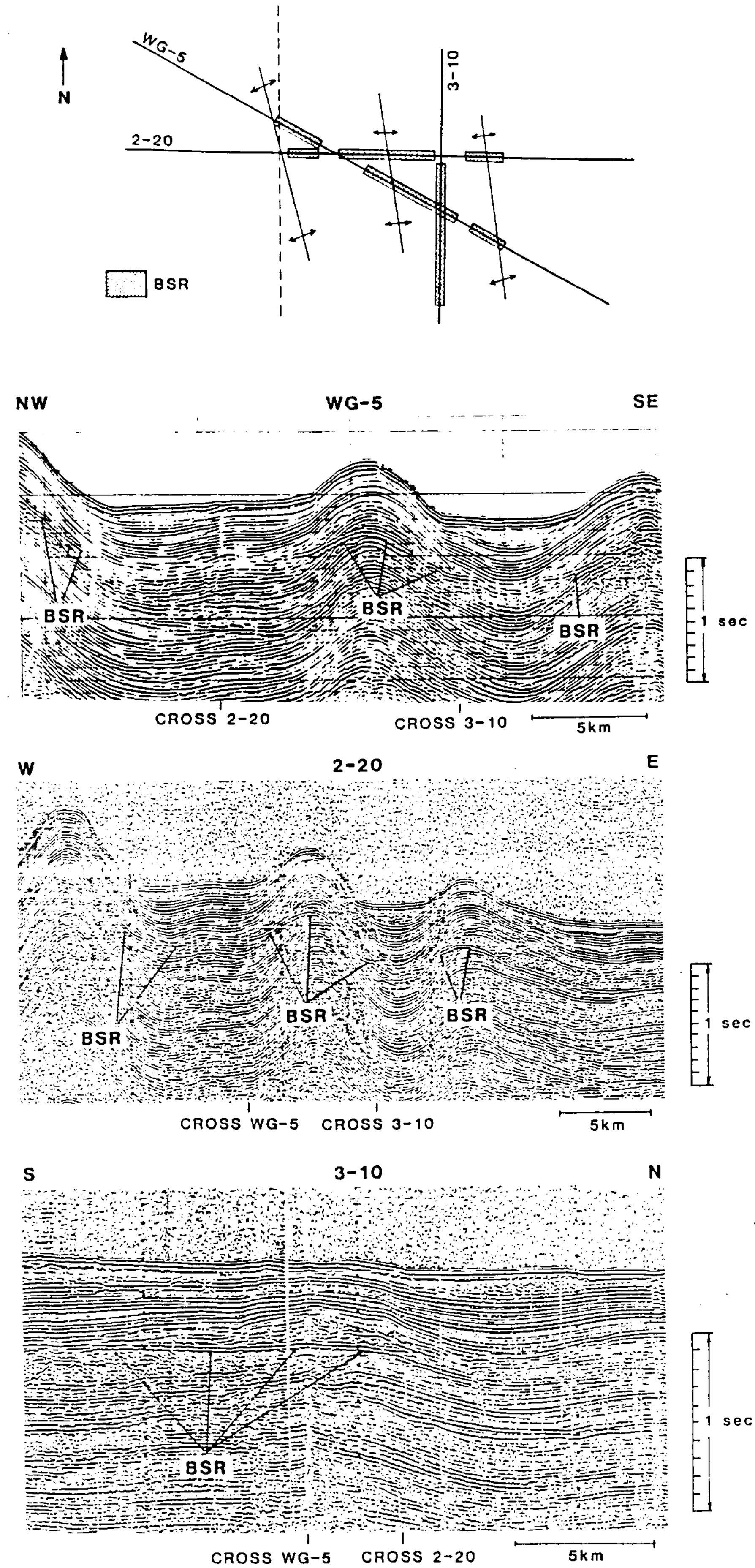


Figure 66. SEISMIC PROFILES WG-5, 2-20, AND 3-10 SHOWING CORRELATION OF BSRs

The BSR can be recognized in both lines 2-20, WG-5, and 3-10 where they intersect (Figure 66). The BSR extends south of line 2-20 on line 3-10 for approximately 8 km, but becomes indistinct less than 1 km north of the intersection. The distance from the sea floor reflector to the top of the first strong reflector of the BSR is 0.47 sec on all three sections.

As illustrated in Figure 67 the previously described reflector at 0.48 sec on line 4-2 is also present in line 3-10. As was the case for line 4-2, the parallel orientation of the sea floor and the sediments prevents the classification of the reflector as a likely hydrate horizon. However, when the entire profile between lines 2-20 and 4-2 along line 3-10 is examined, it can be reasonably inferred that the concordant reflector may be a BSR similar to that found on line 2-20. In Figure 68 the distinct BSR which cuts across the gentle anticline south of line 2-20 becomes diffuse and dies out. A continuous reflector becomes evident at point Q and can be traced south to the intersection with line 4-2. The seismic record is partially obscured by a vertical black stripe 13 km north of the intersection with line 4-2, but a small anticlinal structure can be discerned at 0.2 to 0.4 sec subbottom. Although resolution of the reflectors is difficult beneath the black stripe, the reflector at 0.48 sec subbottom appears to be oblique to adjacent bedding reflectors, similar to the orientation of bedding at the adjacent BSR to the north. If the reflector at the intersection with 4-2 is indeed the same as that under the black stripe, it is likely that the trace at the intersection is a BSR in a concordant sedimentary section, based on long distance correlation with the more pronounced BSR to the north.

The BSR at 2-22 at 0.51 sec subbottom depth correlates with a prominent reflector at 0.49 sec on line 3-10 (Figure 69). It is not apparent whether the difference in travel time indicates that the reflectors are different horizons.

Another north-south seismic line (3-6) intersects the east to west lines which contain BSRs. Line 3-6 is roughly parallel to line 3-10 and approximately 30 km closer to shore. Water depths along line 3-6 average 1,600 m compared with approximately 2,000 m along line 3-10. Line 3-6 intersects BSRs on line 2-22. Although no BSR is found where line 3-6 intersects line 4-2, there is a BSR approximately 5 km east along line 4-2. BSRs are not apparent on line 3-6 where it crosses 2-20 and 4-2, possibly due to concordance of the sea floor and subbottom sedimentary reflectors along the northerly strike of the ridges. Figure 70 illustrates the intersection of lines 2-20 and 3-6. The BSR to the west of the crossing of tracks does not show up in either section at the point of intersection. A reflector which crosses sedimentary horizons is found on line 3-6 at its intersection with 2-22; however, the reflector may be a slump plane rather than a BSR. One possible BSR underlain by a seismic bright spot at 0.48 sec subbottom occurs on line 3-6 between lines 2-20 and 4-2 (Figure 71).

Veracruz Continental Slope

A large number of BSRs were identified offshore of Veracruz south of the main Mexican Ridges fold belt on the continental slope west of the Veracruz Tongue. The Veracruz Tongue is a narrow north-dipping inlet which separates the Mexican continental slope to the west from the salt diapir dominated Campeche Knolls to the east (Figure 1). The BSRs are located beneath 1,600 to 2,400 m of ocean in a broad trend paralleling the Mexican Gulf coast from 80 to 150 km offshore (Figure 55). The sea floor of the Mexican continental slope west of the Veracruz Tongue does not display as

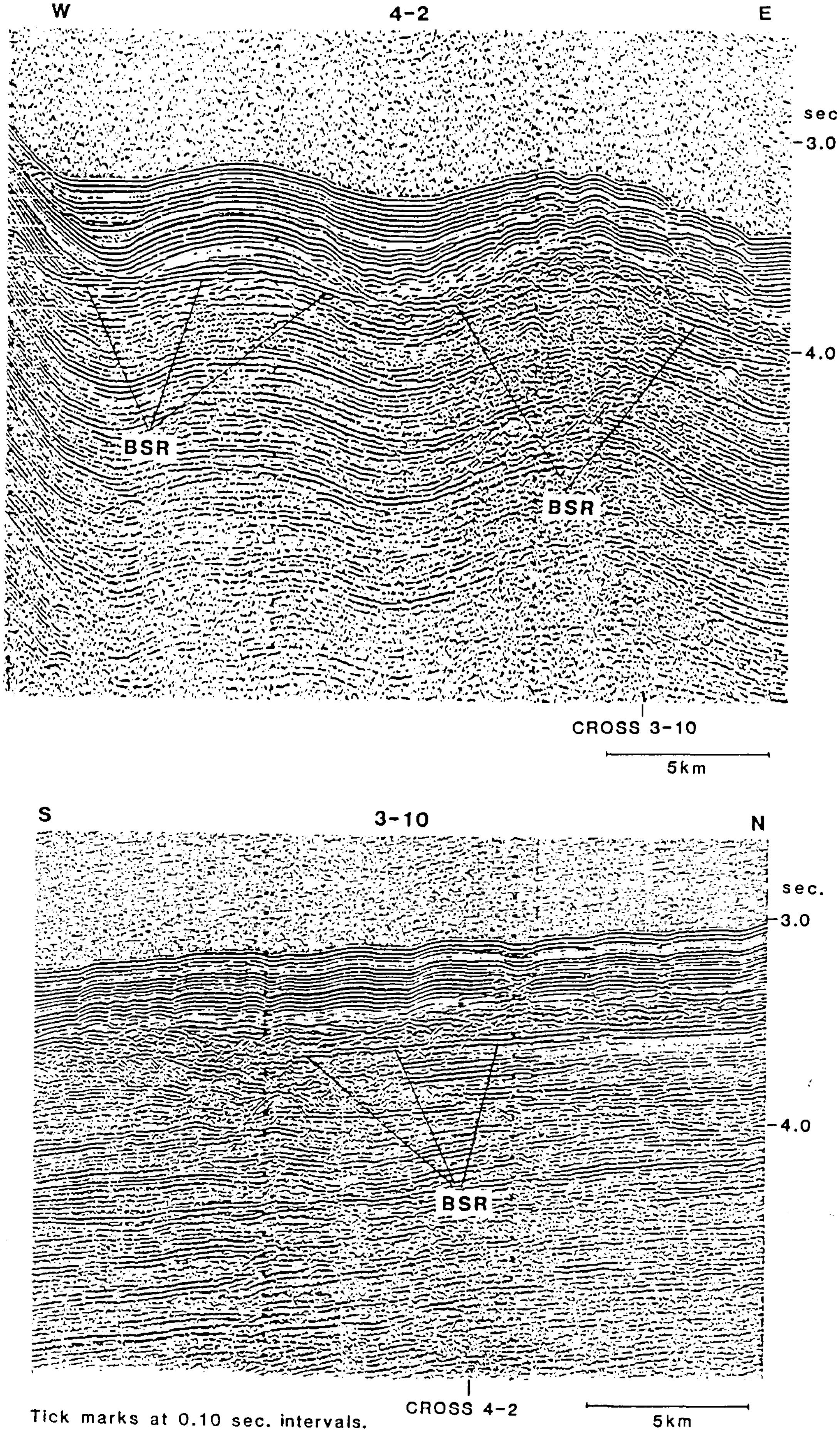
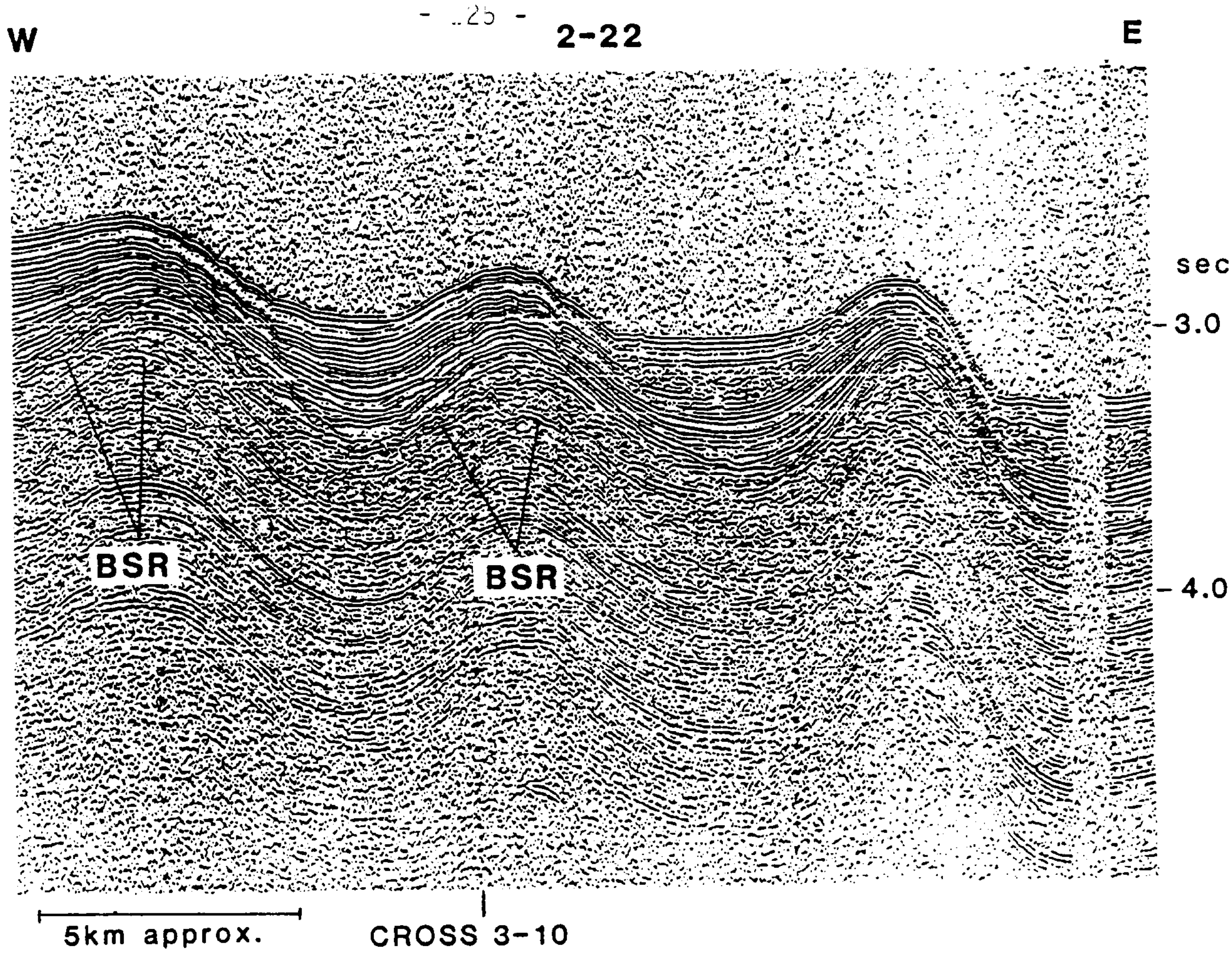


Figure 67. SEISMIC PROFILES 4-2 AND 3-10 SHOWING CORRELATION OF POSSIBLE BSRs

FIGURE 68, Seismic profile 3-10 showing correlation of BSR on line 2-20 with possible BSR on line 4-2, is located in the pocket at the end of the report.



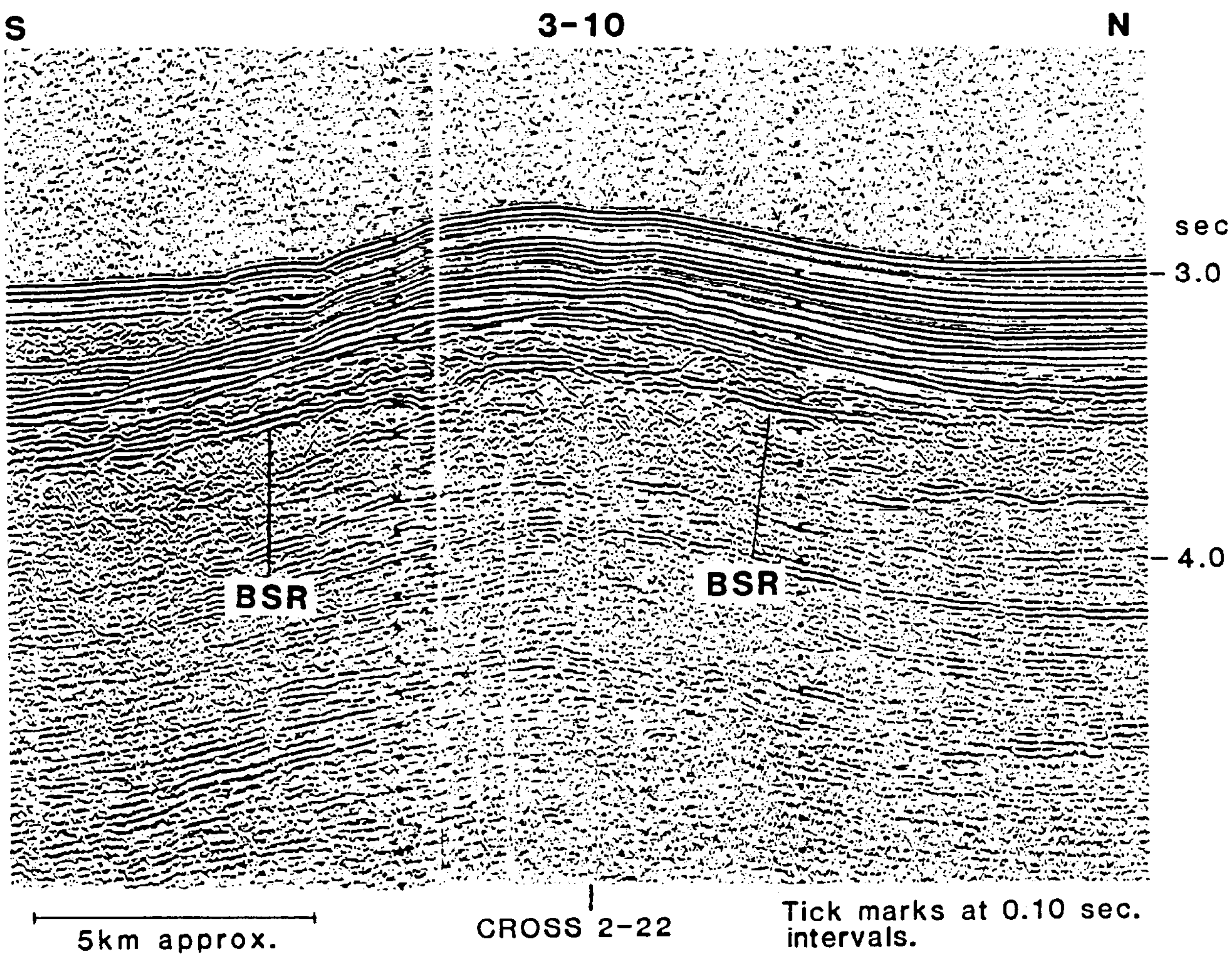
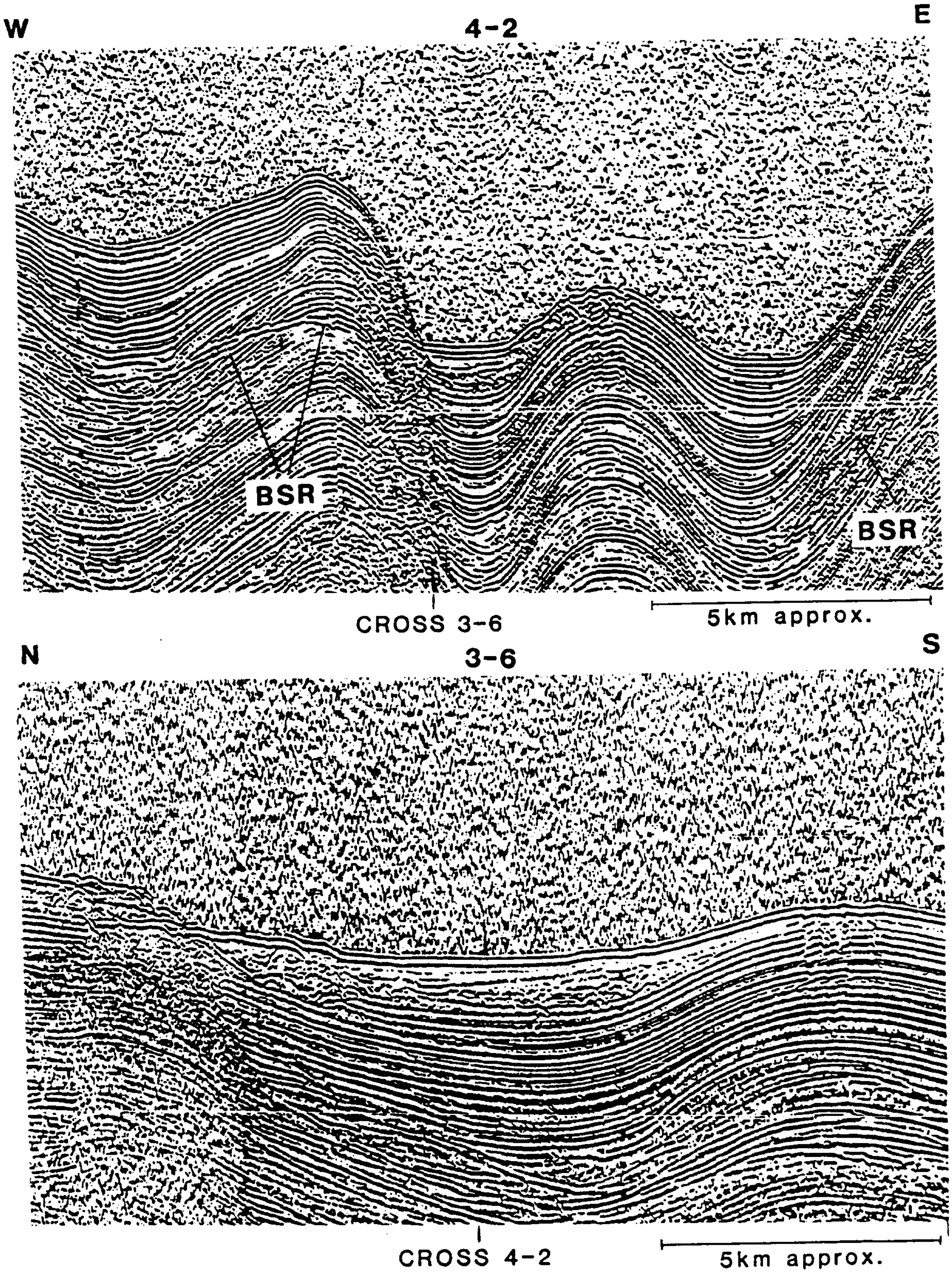
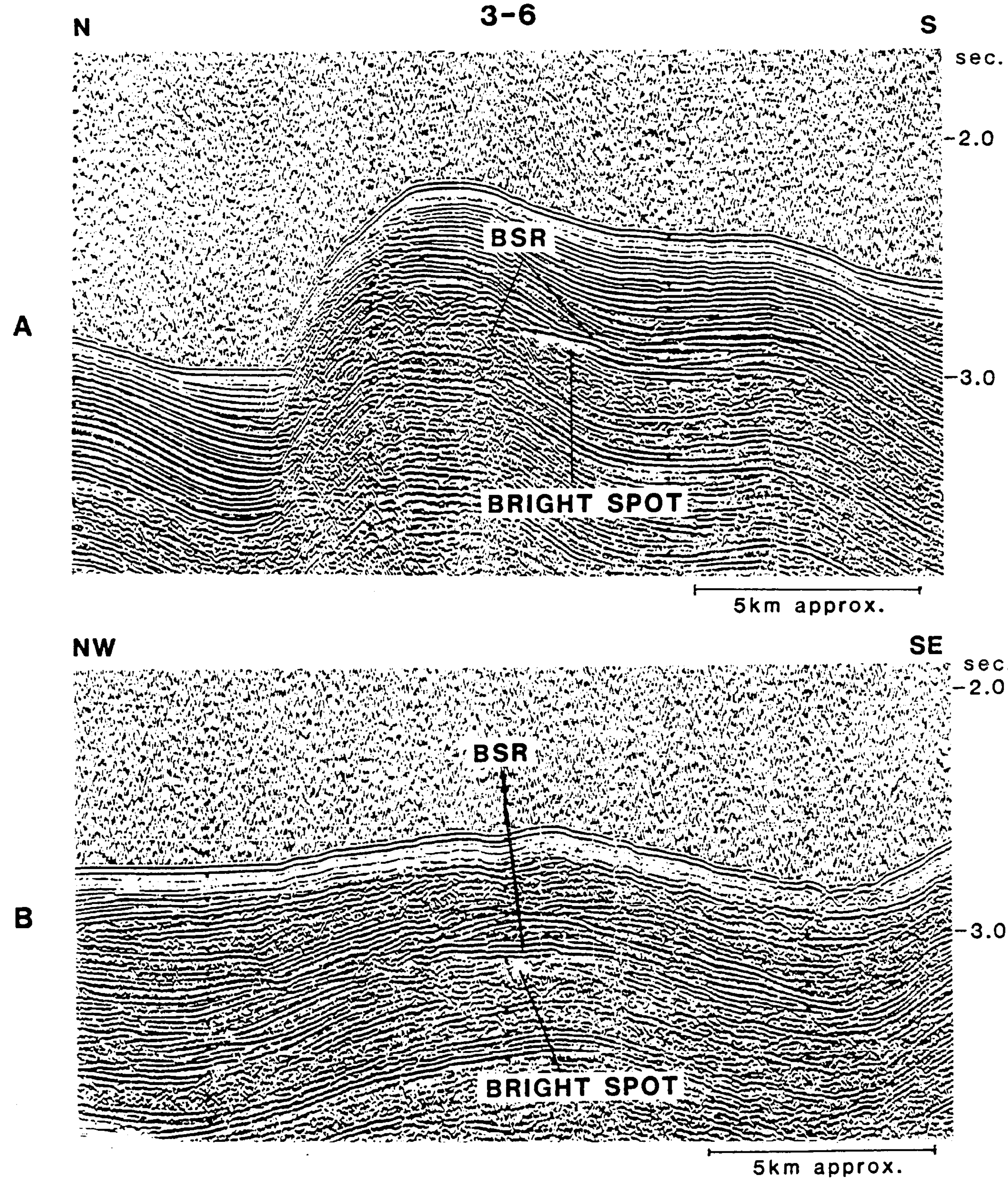


Figure 69. SEISMIC PROFILES 2-22 AND 3-10 SHOWING CORRELATION OF BSRs



Tick marks at 0.10 sec. intervals.

Figure 70. SEISMIC PROFILES 4-2 AND 3-6 SHOWING DISCONTINUITY OF BSRs



A. Located between Lines 4-2 and 2-20. B. Located 110 km N°10 W from A. Tick marks at 0.10 sec. intervals.

Figure 71. BSRs UNDERLAIN BY SEISMIC BRIGHT SPOTS, PROFILE 3-6, SOUTHERN MEXICAN RIDGES

much local relief as the main Mexican Ridges trend to the north, as illustrated by the generally parallel isobaths on Figure 55.

The shale-cored harmonic folds characteristic of the continental slope farther north can be seen, however, in seismic sections offshore of Veracruz. The sea floor topography is more subdued to the south, presumably because folding has been inactive or progressing very slowly for a long enough period of time to permit ponded sediments to fill synclinal depressions, obscuring the folds at the surface. As was the case in the Mexican Ridges proper, BSRs are principally located on the crests or upper flanks of anticlines. The anticlinal locus of the BSRs appears to be more dominant in this southern area than farther to the north. This may be due to the more extensive sedimentary fill of the intervening synclines. The thicker fill in the synclines is generally concordant with the sea floor; only along the anticlinal flanks are discordant beds present at the subbottom depths expected for BSRs (0.4 to 0.6 sec two-way travel time). The BSRs identified were from one east to west profile (2-23), the two north-south profiles described earlier (3-10 and 3-7) from the Kane, and seismic profiles from the 1971 joint USGS - IDOE seismic survey of the Gulf of Campeche. Between 18°N and 21°N, 20 profiles were recorded, 18 of which were approximately east to west. Of these, 14 were found to display pronounced BSRs. For simplicity, the lines with BSRs are referred to by a letter. The relationship between identifying letter and standard date and hour description is summarized in Table 7.

Anticlines found in seismic sections of this area appear to be more continuous than in the Mexican Ridges. This, coupled with the dense spacing of the USGS survey, allows prominent folds to be correlated for distances of 50 km or more. One major anticline can be traced for 40 km through sections C, D, E, F, and G (Figure 72). A prominent BSR is found along the traceable length of the ridge and is most pronounced on the shoreward (west) flank of the structure. The lateral extent of the BSR normal to the strike of the structure (east to west) increases from 3 km on line C through 8 km on line D to 12 km on lines E and F, and diminishes to 7 km on line G. Although gas hydrates may extend far beyond the limits of the BSR into surrounding strata which are concordant with the sea floor, the BSR on this ridge is continuous over an approximately elliptical area of about 400 to 500 km^2. Structural closure beneath the BSR is limited because the eastward dipping sea floor does not reverse slope over the ridge except in lines C and G. East to west closure is approximately 100 m (0.09 sec) on line C and 60 m (0.05 sec) on line D.

A pair of ridges with BSRs can be traced for about 60 km along tracks H, I, 2-23, K, L, M, and N (Figure 73). Little internal structure of the larger shoreward ridge can be determined from the profiles. The chaotic nature of the seismic record in the cores of the anticlines has been interpreted as representing mobilized shale or strata which are too steep to provide discrete reflectors (Garrison and Martin, 1973; Buffler et al., 1979). The smaller ridge has more distinct layers in the cores of the folds. A prominent BSR on the larger anticline can be seen in lines H, 2-23, and I. The BSR covers a north to south distance of 20 km and tapers in width from 6 km at lines H and 2-23 to 3 km at line I, yielding an area of about 100 km^2. About 60 m (0.05 sec) of structural closure can be measured on line 2-23.

The BSR on the smaller seaward ridge covers less distance across the structure than that on the large ridge, but is more persistent along the strike

TABLE 7.

IDENTITY OF USGS BAY OF CAMPECHE SEISMIC LINES

Reference Lable	Cruise Day	Hours GMT
A	154	1000 - 1800
B	155	0800 - 1000
C	156	0600 - 1200
D	150 - 157	2000 - 0900
E	157 - 158	2300 - 0200
F	158	1100 - 1900
G	159	1100 - 1600
H	160	0100 - 0500
I	162	0600 - 1700
K	162	0600 - 1100
L	163	1800 - 2400
M	163	1200 - 1800
N	165	0300 - 0900

FIGURE 72, Seismic profiles across anticline offshore Veracruz, is located in the pocket at the end of the report.

FIGURE 73, Seismic profiles across anticlines offshore Veracruz, is located in the pocket at the end of the report.

of the anticline, being continuous for over 50 km. The width of the BSR across the anticline (east to west) increases to a maximum of 4 km along lines I and K. An approximate area of 180 km^2 is estimated for the BSR. Apparent structural closure of the BSR surface varies from 40 m (0.03 sec) at line I to 100 m (0.08 sec) at lines K and L. The BSR is observed only on the seaward flank of the anticline in section M. If a similar seal exists on the west flank, up to 240 m (0.20 sec) of closure beneath the BSR could result.

Apart from these prominent correlatable ridges, isolated BSRs are found on more landward ridges.

Campeche Knolls

A few isolated BSRs were identified from profiles of the Campeche Knolls, east of the previously described areas. The most readily identified BSRs are from the east flank of the Veracruz Gap where diapiric uplift of the Campeche Knolls salt structures have resulted in an oblique orientation of sediment layers to the sea floor (lines J and E-2; Figure 74). One faint BSR was observed adjacent to a salt structure in the Campeche Knolls. One BSR was found far to the north and east of all others. The reflector was obtained on line 2-17 at 0730 hr in 3,250 m of water. The site is east of the Veracruz Tongue on the south margin of the Sigsbee Abyssal Plain on the flank of a diapiric structure (Figure 75). The two-way travel time of 0.49 sec agrees well with BSRs from other parts of the basin. The short distance over which the BSR can be traced suggests that it may be a small slump deposit from the nearby diapir.

Discussion

BSRs located in this study are restricted to the southwestern part of the Gulf of Mexico. Publicly available seismic sections indicate really extensive BSRs in the southern Mexican Ridges (20.5°N to 22°N) and in the west flank of the Veracruz Tongue and isolated BSRs in and adjacent to the Campeche Knolls. No BSRs were found north of 22°N, although there is evidence of gas hydrates in cores from sites north of 22°N.

The orientation of strata and the sea floor in the areas where BSRs were found allow for recognition of BSRs. However, very similar structures found to the north between 22°N and 25°N display no BSRs. While seismic coverage in the northern Mexican Ridges is not as complete as in the south, it should be statistically sufficient for detection of BSRs if they were to exist in the density found in the south.

The question remains as to whether the northern Mexican Ridges area is barren of gas hydrates, or simply does not display BSRs. Core results suggest that by analogy the northern area should contain hydrates. Gas hydrates were inferred to have been present in cores from DSDP Site 90, on the abyssal plain beneath the northern Mexican Ridges and from Site 89 in a similar setting in the Veracruz Gap seaward of the southern Mexican Ridges.

With no drill hole data from the Ridges themselves, it is difficult to address the question. It is of course possible that the BSRs from the Gulf are merely diagenetic boundaries, although Shipley (1979) claims that BSRs

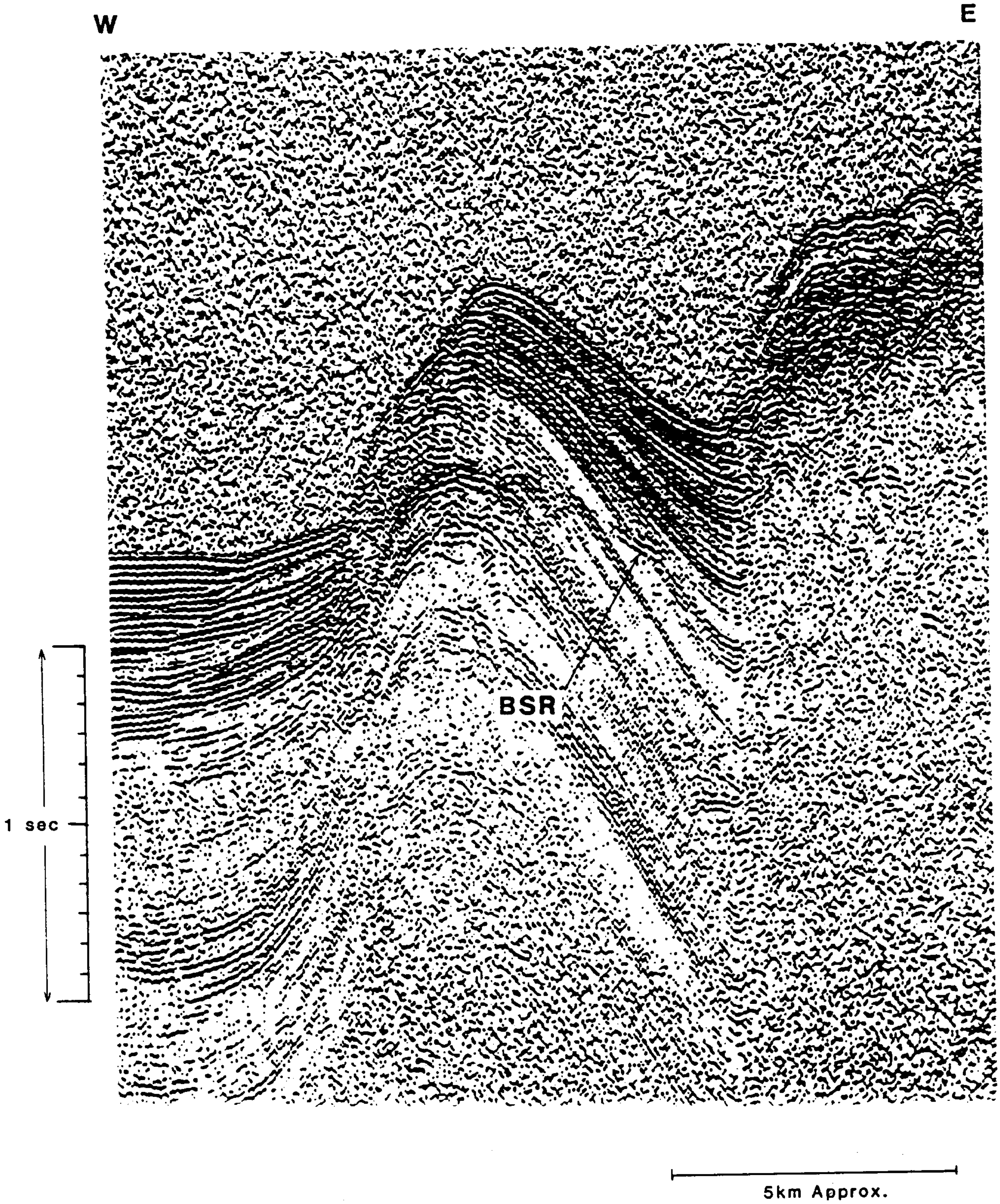
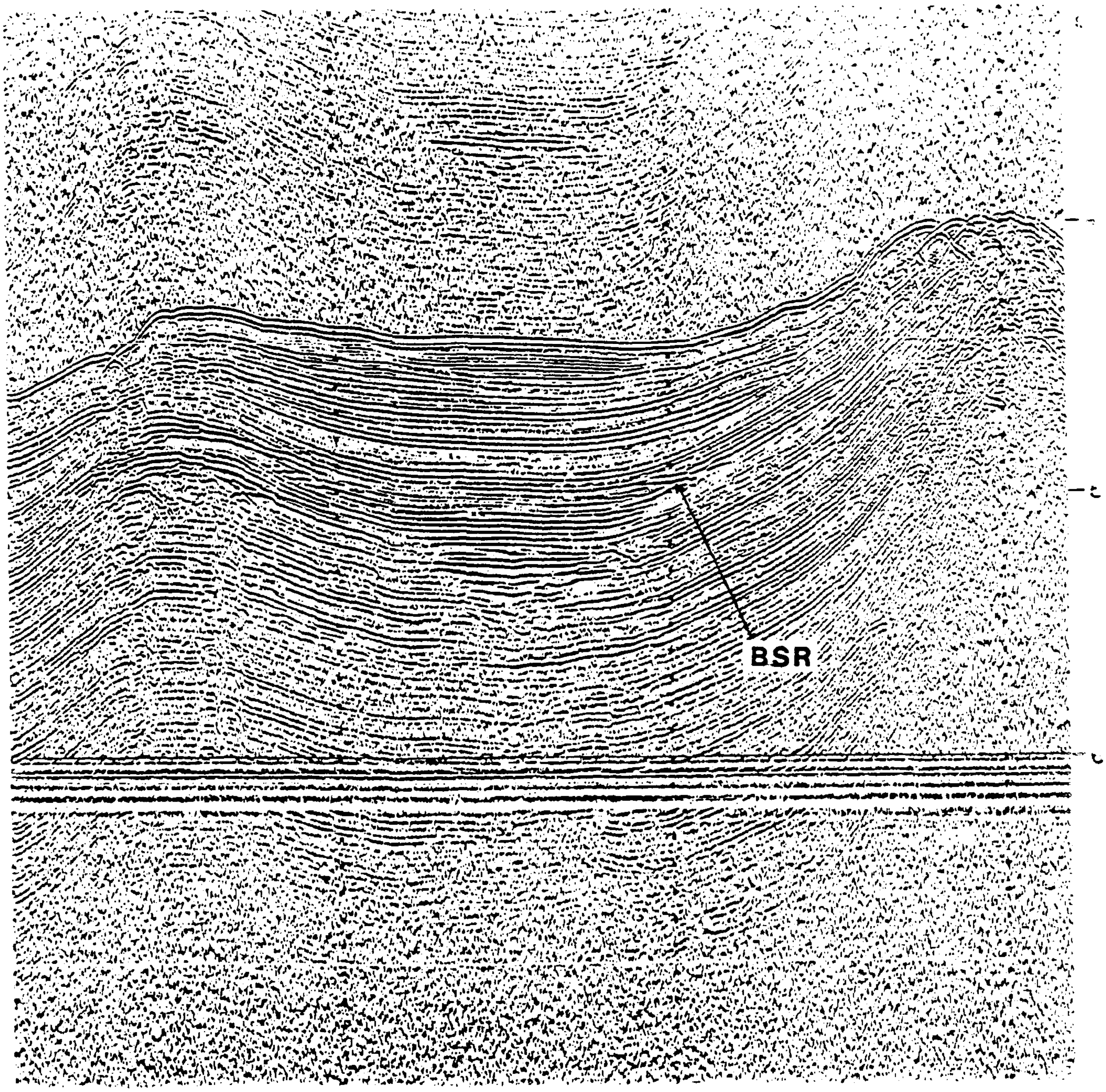


Figure 74. SEISMIC PROFILE I, WEST FLANK OF CAMPECHE KNOLLS



Tick marks at 0.10 sec. intervals.

Figure 75. SEISMIC PROFILE 2-19, NORTH SLOPE OF CAMPECHE KNOLLS, SHOWING POSSIBLE BSR

from the UTIG seismic lines fit the criteria for gas hydrate boundaries. Different sedimentary sources may have resulted in the influx of different amounts of organic matter types with varying gas generating potential which may have affected gas hydrate formation. Alternatively, differences in clastic sediments may mask or enhance seismic reflectivity of the base of the hydrate zone.

One anomaly can be recognized from the scant data available. All four DSDP Sites from the southwest Gulf of Mexico which reported probable gas hydrates (88, 89, 90, 91) also reported varying amounts of volcanic ash in the recovered sediments (Worzel et al., 1973). The ash concentrations were greatest in Miocene sediments. Among the four sites, volcanic ash was much more abundant to the south and west, with discrete layers of ash preserved in the farthest southwest hole, Site 89. This suggested a volcanic source in southern Mexico or Central America (Worzel et al., 1973). Site 89 is immediately downslope from the area with extensive BSRs.

Although the correspondence of BSRs and volcanic ash is probably coincidental, substantial amounts of ash may affect either the potential for gas hydrate formation or the seismic response of a sediment. Ash devitrifies in an aqueous environment to a mixture of clays. Krason and Ridley (1985a) observed that structured water adsorbed on clay may form hydrates at less severe pressure and temperature conditions than bulk water. A band of partially decomposed volcanic ash may thus present an ideal site for nucleation and initiation of gas hydrate crystallization. Volcanic debris also decomposes into varying amounts of zeolites. Zeolites are widely used industrially as catalysts for a variety of processes. It is conceivable that zeolites in sediments may orient water and/or gas molecules in such a way as to substantially lower the energy barriers to formation of gas hydrates. Experience in the Caribbean Sea has shown that the character and intensity of seismic reflections change proximal to volcanic centers (T. Holcomb, personal commun., 1985). Thus, the large proportion of volcanic ash in the sediments of the southwestern Gulf of Mexico may increase the reflectivity of gas hydrate horizons resulting in conspicuous BSRs where the orientation of the strata and sea floor are optimum for BSR recognition. It is interesting to note that reports from DSDP Leg 67, during which massive gas hydrates were cored, mention that the hydrates tended to be localized in thick granular ash layers, similar to those inferred to exist offshore of Veracruz (Hesse and Harrison, 1981).

Assessment of Gas Reserves in Gas Hydrates

Central Gulf of Mexico

The flat lying sediments beneath the Sigsbee Abyssal Plain have been demonstrated by DSDP drilling to contain gas hydrates. The only holes yet drilled in the abyssal environment of the Gulf of Mexico, Sites 90 and 91 displayed degassing characteristics associated with gas hydrates. Of these two sites, degassing was most pronounced at the more westward location, Site 91. The inferred gas hydrates existed from a depth of approximately 150 m subbottom to 838 m at Site 91 and at least 767 m deep at Site 90. A hydrate zone thickness of 600 to 750 m is thus indicated for sediments under the western portion of the Sigsbee Abyssal Plain. There is no evidence of free gas pooled beneath the hydrate layer, however there is no evidence in published reports of Leg 10 that would contradict such a possibility.

No BSRs were located in the Sigsbee Abyssal Plain due to the concordance of strata and the sea floor. Although both holes drilled in the Sigsbee Abyssal Plain encountered evidence of gas hydrates, it cannot be safely assumed that the entire floor of the plain is underlain by hydrates. The shipboard party of Leg 10 reported that Site 91 near the center of the basin exhibited only a fraction of the degassing observed from Site 90 near the western edge of the Sigsbee Abyssal Plain. It would thus be reasonable to assume that Site 91 represents nearly the eastward limit of gas hydrate development in the Sigsbee Abyssal Plain. Similarly, it would be unlikely that gas hydrate deposits are continuous over the entire area west of Site 91. A plausible estimate is that 30% of the abyssal plain between Site 91 and the Mexican continental rise is underlain by a hydrate zone. The concentration of the hydrates in the pore spaces is likewise speculative. Sediments in the depths of interest range from 30% to 50% porosity. To maintain consistency with earlier reports (Krason and Ridley, 1985a, 1985b) it is assumed that 12% of the pore space of the hydrate zone is occupied by gas hydrates, giving a volume fraction of the sediment occupied by gas hydrates of 5%. Using the volume conversion factor of Kuuskraa et al. (1983) for gas hydrates to methane at standard conditions of 1:200, the following relationship will give net amount of methane per volume of hydrated sediment using the above assumptions.

$$\text{volume methane/volume of sediment} = 5\% \times 200 = 10$$

The area under consideration is approximately 50,000 km^2 ($5 \times 10^{10} m^2$) of which 30% is underlain by hydrate. Volume of methane per meter of hydrated sediment over the Sigsbee Abyssal Plain is thus:

$$1 \text{ m depth} \times 5 \times 10^{10} m^2 \text{ area} \times 30\% \text{ areal extent} \times 10 = 1.5 \times 10^{11} m^3 (5 \text{ TCF}).$$

For any given thickness of gas hydrate filled sediment conforming with these assumptions the total methane contained is:

$$\begin{aligned} 1 \text{ m} &= 1.5 \times 10^{11} \text{ m}^3 \text{ (5 TCF)} \\ 10 \text{ m} &= 1.5 \times 10^{12} \text{ m}^3 \text{ (50 TCF)} \\ 100 \text{ m} &= 1.5 \times 10^{13} \text{ m}^3 \text{ (500 TCF)} \\ 500 \text{ m} &= 7.5 \times 10^{13} \text{ m}^3 \text{ (2,500 TCF)} \\ 600 \text{ m} &= 9.0 \times 10^{13} \text{ m}^3 \text{ (3,000 TCF)} \end{aligned}$$

Northwestern Margin of the Gulf of Mexico

Data are scant on the areal distribution of gas hydrates on the continental slope offshore of Louisiana. Although gas hydrates have been recovered from at least eight locations in this area, no published information is available on the proportion of cores containing hydrates to those which were barren of hydrates. Proprietary industry data based on high resolution seismic profiling indicates that Green Canyon Block 184, from which thermogenic gas hydrates were recovered, is underlain by hydrates over approximately 30% to 40% of its area. DSDP Leg 96 cored hydrates in the sediments of the Orca Basin, but all other holes drilled on the cruise were devoid of gas hydrates. These other sites were concentrated on the deeper reaches of the Mississippi Trough and Fan. Considering the good coverage of drill holes across this large alluvial feature and that the shipboard scientific party was specifically searching for signs of hydrates, the gas hydrate potential of the Mississippi Fan appears limited. Biogenic gas hydrates were reported from the Mississippi Canyon lease area (Brooks and Bryant, 1985a) but no details were released on location. Brooks et al. (1985) claims that gas hydrates have been proven to exist over an area of at least 25,000 km^2; however, no estimate of degree of continuity over that area was made. Intraslope basins such as those in the Green Canyon block extend westward approximately 300 km. Approximately 60,000 km^2 of ocean floor deeper than 500 m with structure similar to that of the Green Canyon and Garden Banks areas exists between 90°W and 95°W shoreward of the Sigsbee Escarpment (Figure 1). Since the thermogenic hydrate recovery sites were associated with large active oil seeps, which are less common than salt structures, it seems likely that the 30% areal extent of gas hydrates believed to exist at Block 184 may be anomalously large. An area-wide fraction of sea floor area which is underlain by gas hydrates of 10% appears a more plausible estimate.

Only from the Orca Basin are data on the vertical extent of hydrates available. Gas hydrates were recovered from 26 to 47 m depths with abundant free gas beneath the gas hydrate layer (Bouma et al., 1985). The remainder of the hydrate occurrences in this area for which information is available were obtained with a piston core device with a subbottom penetration of 2 to 3 m. Therefore, maximum depth of occurrence information was not obtained, but gas hydrates sampled with piston coring equipment are present at much shallower depths than are the hydrates in the Orca Basin. Pressure and temperature limits indicate that for the 1,500 m mean depth of the 60,000 km^2 area considered, the maximum thickness of gas hydrates is 300 to 400 m.

Degree of pore filling of some recovered gas hydrates is quite high. Nodular chunks and disseminated crystals have been reported. Once again lack of pertinent data on the cores precludes quantification of the degree of pore filling by the hydrates. In absence of hard data to the contrary, we shall continue to use 5% as the estimate of volume fraction of gas hydrates in sediment.

In the northwest continental margin, the volume of gas in hydrate form per meter thickness is:

1 m depth x 6 x 10^{10} m^2 area x 5% hydrates x 200 gas/hydrate conversion factor x 10% areal extent of hydrate layer = 6 x 10^{10} m^3 (2 TCF)

Gas volumes for any chosen area wide mean gas hydrate zone thickness are:

1 m = 6 x 10^{10} m^3 (2 TCF)
10 m = 6 x 10^{11} m^3 (20 TCF)
100 m = 6 x 10^{12} m^3 (200 TCF)
300 m = 1.8 x 10^{13} m^3 (600 TCF)

Western Margin of the Gulf of Mexico

Of the three physiographic areas of the western Gulf of Mexico studied in this report, only the southwestern area comprising the Mexican continental slope, the Veracruz Tongue, and the Campeche Knolls displayed BSRs. As such, a more confident estimate of potential gas hydrate resources is possible. Since it has been reported that BSRs are caused or enhanced by free gas trapped beneath the impermeable hydrate seal (Bryan, 1974), this area with well developed BSRs may also contain substantial free gas reserves.

Areas with Bottom Simulating Reflectors. Four particularly distinct ridges underlain by BSRs were correlated throughout their length with seismic profiles (Figures 55, 63, 72, 73). BSRs under these four anticlinal ridges alone cover over 1,250 km^2. Some 70% of the BSRs observed in this area are in locations other than these ridges. If the areal extent of these isolated BSRs is similar to the correlated ridges, then it is likely that closely spaced seismic surveys would reveal that over 5,000 km^2 of BSRs exist. It should be noted that the seismic data was not collected or processed so as to maximize BSR resolution. The area underlain by BSR may be much greater than 5,000 km^2; improved seismic methods may detect BSRs in areas of only slight discordance of strata and sea floor and may allow tracing of the BSRs farther down the flanks of anticlines, greatly increasing the cumulative area of ocean bottom underlain by prominent BSRs. As previously discussed, BSRs are limited to areas south of 22°N and are concentrated west of 95°W. One reflector with the characteris- tics of a BSR, but which was much too faint for confident assignment as a BSR was noted on UTIG line WG-3W at 23.8°N, 80 km upslope from DSDP Site 90 (Figure 1). Lack of corroborative evidence in nearby seismic lines dictated that it not be included with the much more distinct BSRs clustered to the south.

Using the assumptions of degree of pore filling derived for previous calculations, the volume of methane per meter thickness in areas reasonably inferred to be underlain by BSRs can be computed.

$$1 \text{ m thickness} \times 5 \times 10^{9} \text{ m}^2 \text{ area} \times 5\% \text{ hydrates} \times$$
$$200 \text{ volume methane/volume hydrates} = 5 \times 10^{10} \text{ m}^3 \text{ (1.7 TCF)}$$

Exact depths of BSRs and thus thickness of the hydrate zone is strongly dependent on the seismic velocity assigned to the sediments. In the absence of detailed seismic velocity data for the area, it can be assumed that the BSR typically occurs at 400 to 700 m subbottom. Cores from the nearby abyssal plain indicate that hydrates are probably not present at depths of less than 120 to 150 m, giving a net hydrate zone thickness of 250 to 650 m for a mean of 400 m.

The total gas contained in BSR underlain hydrate layers of different mean thickness would thus be:

$$1 \text{ m} = 5 \times 10^{10} \text{ m}^3 \text{ (1.7 TCF)}$$
$$10 \text{ m} = 5 \times 10^{11} \text{ m}^3 \text{ (17 TCF)}$$
$$100 \text{ m} = 5 \times 10^{12} \text{ m}^3 \text{ (170 TCF)}$$
$$250 \text{ m} = 1.2 \times 10^{13} \text{ m}^3 \text{ (420 TCF)}$$
$$400 \text{ m} = 2 \times 10^{13} \text{ m}^3 \text{ (680 TCF)}$$
$$650 \text{ m} = 3 \times 10^{13} \text{ m}^3 \text{ (1100 TCF)}$$

Campeche Knolls. Evidence from DSDP Leg 10 indicates that gas hydrates are present in the southwestern areas of the study region which do not show pronounced BSRs. None of the previously described gas hydrate occurrences offshore of Louisiana were associated with BSRs (W.R. Bryant, personal commun.). Since BSRs may be a result of gas trapping beneath sediments, possible gas hydrate deposits in areas without pronounced BSRs in the southwestern portion of the Gulf of Mexico are here evaluated.

DSDP Site 88 drilled on a diapir in the Campeche Knolls (Figure 46), probably contained gas hydrates in the sediments above the diapir. Similar diapirs which may also contain gas hydrates abound throughout the Campeche Knolls. Intraslope basins analogous to those in the Green Canyon area from which gas hydrates were recovered are also common in the Campeche Knolls. Approximately 25,000 km^2 of the Campeche Knolls have depth and water temperature conditions favorable for gas hydrate formation. Without more evidence for actual gas hydrate presence than five BSRs (A2, B2, J, E, 2-17) and the results of DSDP Site 88 in the Campeche Knolls, a lower estimate of lateral continuity may be indicated; 5% of the area of the Campeche Knolls being a host for gas hydrates is the selected extent.

The amount of gas per meter depth of gas hydrated sediment is thus:

1 m depth x 2.5 x $10^{10} m^2$ area x 5% hydrates x 200 volume of gas/volume of hydrate x 5% areal extent = 1.25 x $10^{10} m^3$ (0.4 TCF)

BSRs on the deep flanks of the Campeche Knolls area are typically at depths of 500 to 600 m subbottom. Drilling at Site 88 was terminated before reaching the bottom of the suspected hydrate zone. However, little or no gas was experienced at 100 m (Worzel et al., 1973). The projected typical gas hydrate zone thickness at the 1,500 m typical depth based on this sparse information is 300 m.

Gas contained in hydrates in the Campeche Knolls at various thicknesses of hydrate layers is:

1 m = 1.2 x $10^{10} m^3$ (0.4 TCF)
10 m = 1.2 x $10^{11} m^3$ (4 TCF)
100 m = 1.2 x $10^{12} m^3$ (40 TCF)
300 m = 4 x $10^{12} m^3$ (120 TCF)

Veracruz Tongue. The nearly flat sediments of the lower continental rise beneath the Veracruz Tongue were shown to contain large amounts of gas in volcanic ash-rich sediments. At DSDP Site 89, cores presumably containing gas hydrates were recovered 60 km downslope (east) of previously described BSRs. Approximately 15,000 km^2 of sediments similar in depositional style and presumably lithologic type are found under 2,500 to 3,500 m of water in continental rise and lower slope environments. BSRs are not found in this area due to the general concordance of sedimentary layers and sea floor. Distinct, high amplitude reflections are found throughout the sediments under the Veracruz Tongue at reflection times consistent with those expected of BSRs (Figure 65). Although the one drill hole (DSDP Site 89) which penetrated to the minimum subbottom depth of hydrate occurrence in the southwestern Gulf of Mexico area (approximately 120 m), certainly does not constitute a statistically significant sample, it is interesting that gas hydrates were inferred to have been cored there.

The thickness of the probable hydrate layer at Site 89 was at least 320 m (Figure 54). Using geothermal gradients reported by Epp et al. (1970) the hydrate stability zone could be as thick as 500 to 600 m.

With no information on the areal distribution of gas hydrates beneath the Veracruz Tongue it is proposed to use the same figure previously used for the Sigsbee Abyssal Plain (30%). The volume of gas at standard conditions per meter of thickness of gas hydrated sediment is:

$$1 \text{ m thickness} \times 1.5 \times 10^{10} \text{ m}^2 \text{ area} \times 5\% \text{ hydrate} \times 200 \text{ volume gas/volume hydrate} \times 30\% \text{ areal extent of hydrates} = 4.5 \times 10^{10} \text{ m}^3 \text{ (1.5 TCF)}$$

The volumes of gas contained in gas hydrates beneath the Veracruz Tongue for various area-wide thicknesses of hydrated sediments are:

1 m =	4.5×10^{10} m^3 (1.5 TCF)
10 m =	4.5×10^{11} m^3 (15 TCF)
100 m =	4.5×10^{12} m^3 (150 TCF)
300 m =	1.3×10^{13} m^3 (400 TCF)
500 m =	2×10^{13} m^3 (600 TCF)

Northern Mexican Ridges. North of the area of the continental slope displaying BSRs (19°N to 22°N) a large area with very similar geomorphological, structural, and sedimentological conditions exists. The absence of BSRs in this area was previously discussed and attributed to possibly less reflective sediments or lack of extensive gas hydrates. The proximity of DSDP Site 90 to this area, 40 km downslope (east), argues that some hydrates should exist in this area. Indeed, the sediments of the northern Mexican Ridges area would be expected to be more highly efficient in methane production than the resedimented turbidites of the abyssal plain whose organic matter had been exposed to oxidative degradation for another cycle.

In the absence of any evidence that gas hydrates do in fact exist in the continental slope sediments of the northern Mexican Ridges, an estimate of the areal extent of gas hydrates of only 2% is proposed for the 40,000 km^2 area.

The volume of methane in dispersed gas hydrates in the Mexican Ridges per meter of hydrated sediments is 1:

$$\text{m thickness} \times 4.0 \times 10^{10} \text{ m}^2 \text{ area} \times 5\% \text{ hydrates} \times 200 \text{ volume gas/volume hydrates} \times 2\% \text{ areal extent} = 8 \times 10^{9} \text{ m}^3 \text{ (0.3 TCF).}$$

Gas volumes from gas hydrates in the northern Mexican Ridges for various area-wide depths of thicknesses of hydrate sediment layers is:

1 m =	8×10^{9} m^3 (0.3 TCF)
10 m =	8×10^{10} m^3 (3 TCF)
100 m =	8×10^{11} m^3 (30 TCF)
300 m =	2.4×10^{12} m^3 (90 TCF)
500 m =	6×10^{12} m^3 (150 TCF)

Southern Mexican Ridges and Veracruz Continental Slope. Sediments sufficiently deep to stabilize gas hydrates exist in areas of the southern Mexican Ridges and Veracruz continental slope which are not underlain by BSRs. It is highly likely that gas hydrates exist in some sediments which are adjacent to BSR bearing sediments. The interpretation of BSRs as representing only gas hydrate zones underlain by free trapped gas (Bryan, 1974), and the identification of gas hydrates at numerous locations in the Gulf of Mexico in the absence of BSRs further suggests that sediments adjacent to areas underlain by BSRs contain gas hydrates.

Of the 35,000 km^2 area, 5,000 km^2 underlain by BSRs has already been assessed. The remaining 30,000 km^2 contains a wide range of sediments in a variety of physical settings. A conservative estimate of the fraction of the area exclusive of the BSR zones which would contain gas hydrates is 15%.

Thicknesses of these purported gas hydrate deposits may range up to the thickness of nearby BSR-documented zones, 500 m. Average thickness of the widely scattered hydrate deposits without BSRs is estimated to be closer to 200 m.

The volume of gas contained in gas hydrate sediments without BSRs in the southern Mexican Ridges and Veracruz continental slope areas per meter of thickness of 5% hydrated sediment is:

$$\begin{gathered}1 \text{ m thickness} \times 9 \times 10^{10} \text{ m}^3 \text{ area} \times 5\% \text{ hydrates} \times \\ 200 \text{ volume gas/volume hydrates} \times 15\% \text{ areal extent} = \\ 1.3 \times 10^{11} \text{ m}^3 \text{ (5 TCF)}\end{gathered}$$

1 m =	1.3×10^{11} m^3 (5 TCF)
10 m =	1.3×10^{12} m^3 (50 TCF)
100 m =	1.3×10^{13} m^3 (500 TCF)
200 m =	2.6×10^{13} m^3 (1000 TCF)

Assessment of Gas Reserves Beneath Gas Hydrates

In view of the technical difficulties associated with producing gas by dissociation of natural gas hydrates, gas trapped beneath an impermeable gas hydrate seal may represent the most economically recoverable energy resource of hydrate filled sediments. The high amplitude, persistent BSRs in the anticlines of the southern Mexican Ridges and Veracruz continental slope present the possibility of trapping tremendous amounts of gas in structural highs beneath gas hydrate seals.

Evaluating the amount of structural closure and thus the volume of gas potentially trapped is complicated by lack of certainty of continuity of the hydrate zone and the permeability of the reservoir sediments in the cores to migration along the strike of the ridge. Although as much as 840 m of closure was noted beneath a BSR across a large anticline, pooling of a correspondingly large column of gas would be limited by breaks in gas hydrate layer continuity along the strike which would provide a spillpoint for gas to would be escaping and hydrates would form, establishing a perfect seal. Although bathymetry of the area (Figure 55) shows that some anticlinal ridges plunge at both their north and south ends, the overall northeasterly regional dip results in most anticlines geing closed by plunge only on their northern ends. Most of the folds merge into the gradual slope of the continental margin at their southern termini producing north plunging noses rather than doubly plunging anticlines. These noses (singly plunging anticlines) are open to southerly migration of hydrate trapped gas which could then spill from the trap. If the permeability of the reservoir sediments in the anticlines is limited, the gas may not drain by southerly migration from the anticline, and may accumulate in a combination structural-stratigraphic trap. No estimates of the permeability of the Mexican Ridges sediments at 600 m depths are available, but an anticipated porosity of approximately 40% and the generally semiconsolidated nature of the sediments at comparable depths in DSDP Leg 10 indicates that such long-range (10 - 80 km) migration of gas resulting in draining of large amounts of gas from beneath gas hydrate zones in the anticlines is not entirely unreasonable.

Estimates obtained earlier of BSR closure normal to the strike of the anticlines averages 20 to 60 m. If lateral drainage is significant, the gas may be trapped only in thin zones immediately beneath the hydrate layer, if at all.

Estimates of gas in place beneath the gas hydrate seal can be obtained for various area-wide mean heights of the gas column by making reasonable assumptions of reservoir parameters. The typical bathymetric depth of the anticlines underlain by BSRs is 2,000 m. The 600 m subbottom depth of BSRs indicates that the gas pools will be trapped at a typical depth of 2,600 m below sea level. If hydrostatic conditions can be assumed for the sedimentary pile, the sub-hydrate reservoir would be under approximately 260 atm of pressure; gas volumes at reservoir pressures should be multiplied by 260 to convert to standard conditions (1,000 Btu/ft.^{3}). If the water saturation of such a reservoir were 60% and porosity were 40% gas would occupy 16% of the rock volume.

The volume of gas per km^2 of reservoir 1 m thick would be:

1 m thickness x 1 x 10^6 m^2/km^2 area x 16% gas volume fraction x 260 gas expansion factor = 4 x 10^7 m^3 (1 MMM CF)

Extending the per km^2 figure to the approximately 5,000 km^2 areal extent of BSRs in the southwestern Gulf of Mexico results in 2 x 10^{11} m^3 (7 TCF) of pooled gas at standard conditions per meter of reservoir thickness.

Area-wide totals for gas trapped beneath gas hydrate barriers for various mean heights for the gas column are:

1 m =	2 x 10^{11} m^3 (7 TCF)
10 m =	2 x 10^{12} m^3 (70 TCF)
20 m =	1 x 10^{12} m^3 (350 TCF)
50 m =	1 x 10^{13} m^3 (350 TCF)

Conclusion

Possible gas hydrate resources beneath the Western Gulf of Mexico are substantial. Gas hydrates have been cored from shallow to moderate ocean depths (500 to 1,200 m) on the northwestern continental margin. Gas hydrate presence has been inferred in cores from 3,000 to 3,700 m ocean depths in the central and southwestern portions of the study region. Extensive BSRs are found at water depths of 1,500 m to 2,700 m in the southwestern margin of the western Gulf of Mexico region where structural disturbance of the sediments permits their recognition. Extrapolation of geological conditions present at these sites indicates that gas hydrates probably exist in many additional locations throughout the western Gulf of Mexico.

Several factors of the geology of the western Gulf of Mexico appear to be well suited to gas hydrate formation and stabilization. Sediment influx from the Mississippi, Rio Grande and numerous Mexican rivers imparted adequate amounts of terrestrial, gas-prone organic matter to the deeper parts of the Gulf. Rapid sedimentation, especially during the late Tertiary and the Quaternary, enhanced preservation of the organic matter. Structural disturbance in the form of diapirism, growth and thrust faulting, and decollement folding resulted in migrational conduits from deep thermogenic source beds and from zones of efficient microbial methanogenesis to the cool sediments of the gas hydrate stability zone.

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